PETROGRAPHY, PETROCHEMISTRY AND PETROGENESIS OF HURONIAN VOLCANIC ROCKS OF THE ELLIOT LAKE REGION, ONTARIO

A Thesis Presented to the
Department of Geological Sciences
Brock University

In Partial Fulfillment
of the Requirements for the Degree
Master of Science
in Geological Sciences

by
Bradley Cumming

Dr. W. T. Jolly — Thesis Supervisor

TABLE OF CONTENTS

page

Abstract 1

Acknowledgements 2

Introduction 3

Previous work 6

General Geology 9

Regional Geology 14

Petrography 23

Metamorphism 32

Petrochemistry 36

Trace Element Chemistry 46

Discrimination Diagrams 50

Discussion 54

Conclusions 75

References 78

Appendix 89

LIST OF FIGURES

Page

1. Location of the Elliot Lake area 4
2. General geology map 5
3. Thessalon Formation volcanic rocks 15
4. Volcanics with quartz veins and stringers 17
5. Small scale brecciation in volcanics 17
6. Plagioclase glomerophenocrysts 18
7. Vugs occurring in close association with amygdules 18
8. Plagioclase glomerophenocrysts 21
9. Volcanics illustrating schistosity and vugs 21
10. Temperature and pressure conditions of Greenschist metamorphism acting on Elliot Lake volcanic rocks 33
11. AFM diagram 37
12. Variation diagrams of MgO versus major oxides 39
13. FeO + Fe_2O_3 - SiO_2 diagram 40
14. FeO / MgO - FeO diagram 41
15. FeO / MgO - TiO_2 diagram 43
16. Variation diagrams of MgO versus trace element abundances 47
17. Variation diagrams of MgO versus trace element abundances 48
18. Zr - TiO_2 *100 - Ni diagram 52
19. Zr - TiO_2 / 100 - Y*3 diagram 53
20. ACF diagram 58
21. FeO / MgO - FeO scatter diagram 65

22. FeO / MgO - TiO_2 scatter diagram..... 66

23. Regression line for FeO / MgO - FeO scatter diagram..... 67

24. Regression line for FeO / MgO - TiO_2 scatter diagram..... 69

25. Tectonic model illustrating the evolution of Elliot Lake volcanic rocks (modified after McNaughton, 1985)..... 71/72

LIST OF PLATES

page

1. Atoll texture evident in plagioclase microlites..... 24
2. Epidotization of plagioclase..... 24
3. Interstitial texture of Greenschist assemblage minerals..... 25
4. Biotite forming at the edges of chlorite crystals..... 25
5. Epidotization of plagioclase laths..... 27
6. Plagioclase glomerophenocrysts illustrating an atoll texture..... 27
7. Acicular chlorite needles in quartz crystals..... 29
8. Hydrogarnets situated in chlorite amygdules..... 29
9. Actinolite rimming and incising chlorite..... 31
10. Interstitial texture of Greenschist assemblage minerals..... 31

LIST OF TABLES

page

1. Comparison of stratigraphic nomenclature used for north shore of Lake Huron 7
2. Lithology and depositional environment of rocks of the Huronian Supergroup 11
3. Metamorphic mineral assemblage of the study area 35
4. Statistical data for tholeiitic rocks 64

ABSTRACT

In the Elliot Lake region of northern Ontario, volcanic lava piles represent the lowermost units of the Huronian Supergroup. These rocks generally trend east-west and belong to the Elliot Lake Group. They are situated on the north and south limbs of the Quirke Lake Syncline.

The volcanic rocks of this study contain a secondary mineral assemblage consisting of actinolite, biotite, chlorite, epidote/clinozoisite titanomagnetite and calcite characteristic of greenschist metamorphism. Compilation of data suggests that metamorphism of the volcanic rocks proceeded between 325° and 425°C and between 2.4 and 4.7 kb.

Geochemically these lavas represent tholeiitic and calc-alkaline assemblages. The tholeiites are characteristically enriched in Fe and Ti and consist mainly of basalts, basaltic andesites and andesites. These rocks are believed to have formed by the partial melting of a peridotite source at low P-T. In contrast, the calc-alkaline rocks are depleted in Fe and Ti, but show a significant enrichment in Si and Zr; andesite is the major rock type for this assemblage. It is postulated that the calc-alkaline suite of rocks was the result of either the partial melting of a basaltic magma at shallow depth, or the melting of sialic crustal material due to the added weight of tholeiitic material on an unstable crust and to downwarping processes initiated by convection cells.

ACKNOWLEDGEMENTS

I wish to thank Dr. W. T. Jolly for his advice, supervision and suggestions during the course of this study. Discussions with Drs. H. R. Williams and P. A. Peach are also greatly appreciated. Many thanks to Ed Lorek for his assistance during the gathering of field data , to P. Brown and Danny Soo for helping me with drafting. I am also indebted to Joan Morrison for her aid in the preparation of geochemical samples and constant encouragement. I also wish to thank Dr. U. Brand for his assistance during the compilation of statistical data and review of the thesis. My sincerest gratitude is extended to Shirlee Wallace who offered to type parts of this manuscript.

INTRODUCTION

The Elliot Lake region is part of the Precambrian Shield situated within the Huronian Supergroup. Elliot Lake is located approximately 129 km east of Sudbury and west of Sault Ste. Marie (Fig. 1). The map area is located northeast and west of Elliot Lake, Ontario, within Joubin, Bouck, Beange and Nicholas Townships, between latitudes 46°22' and 46°34.7" N and longitudes 82°30' and 82°55' W (Figs. 1, 2). It contains the Pecors Lake, Dollyberry Lake and Crazy Lake volcanics respectively. North of Ten Mile Lake and south of Flack Lake is a belt of Dollyberry Lake volcanics that was not sampled due to its inaccessibility. These volcanic rocks represent the Elliot Lake Group and can be correlated with the Dollyberry Lake rocks collected for this study (Robertson 1967, 1972). Accessibility to the study area was provided both directly and indirectly by Highway 108.

Field work was conducted by the author during May and June of 1985 and 110 samples were collected for petrographic and geochemical analysis.

Volcanic rocks collected from the Elliot Lake region of Ontario belong to the Thessalon Formation. Samples are predominantly basaltic in composition, and basaltic andesites and andesites also occur. The volcanics of the Thessalon Formation represent the lowermost member of the Huronian Supergroup; stratigraphic reconstructions suggest they were overlain by up to 12.2 km of sediments, yielding a total pressure during alteration of 2.4 to 4.7 kb.

Little attention has been directed towards the metamorphic and chemical history of these volcanic rocks, and the purpose of this study is to present a petrographic, geochemical and petrogenetic model for the Thessalon Formation volcanic rocks of the Elliot Lake region of northern Ontario.

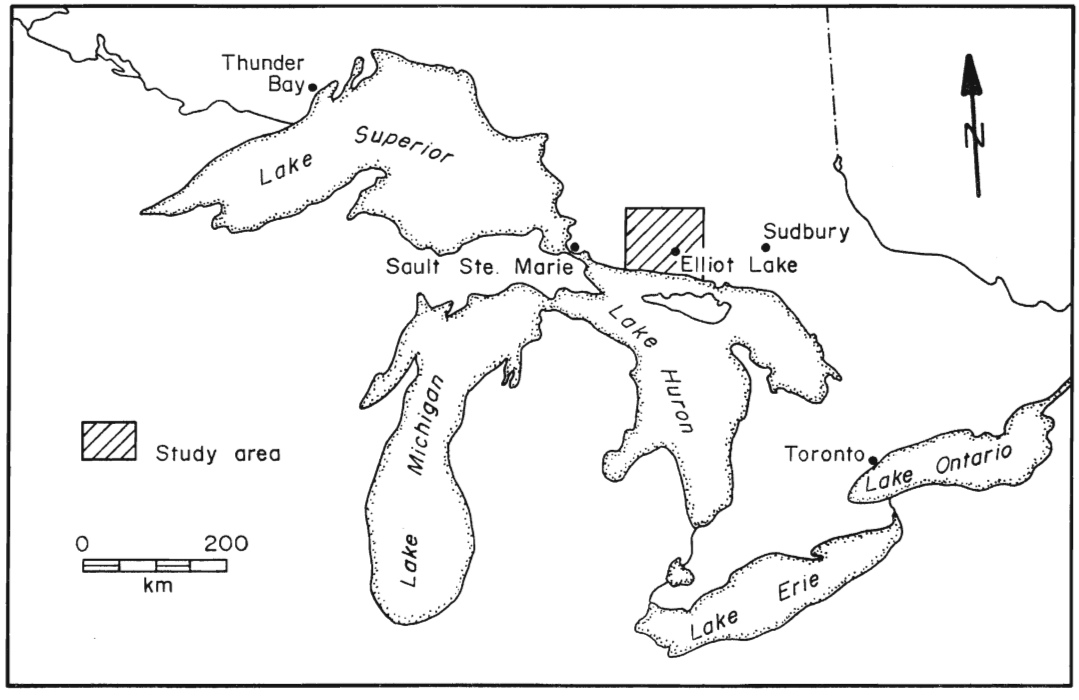

Figure 1. Location of the Elliot Lake study area.

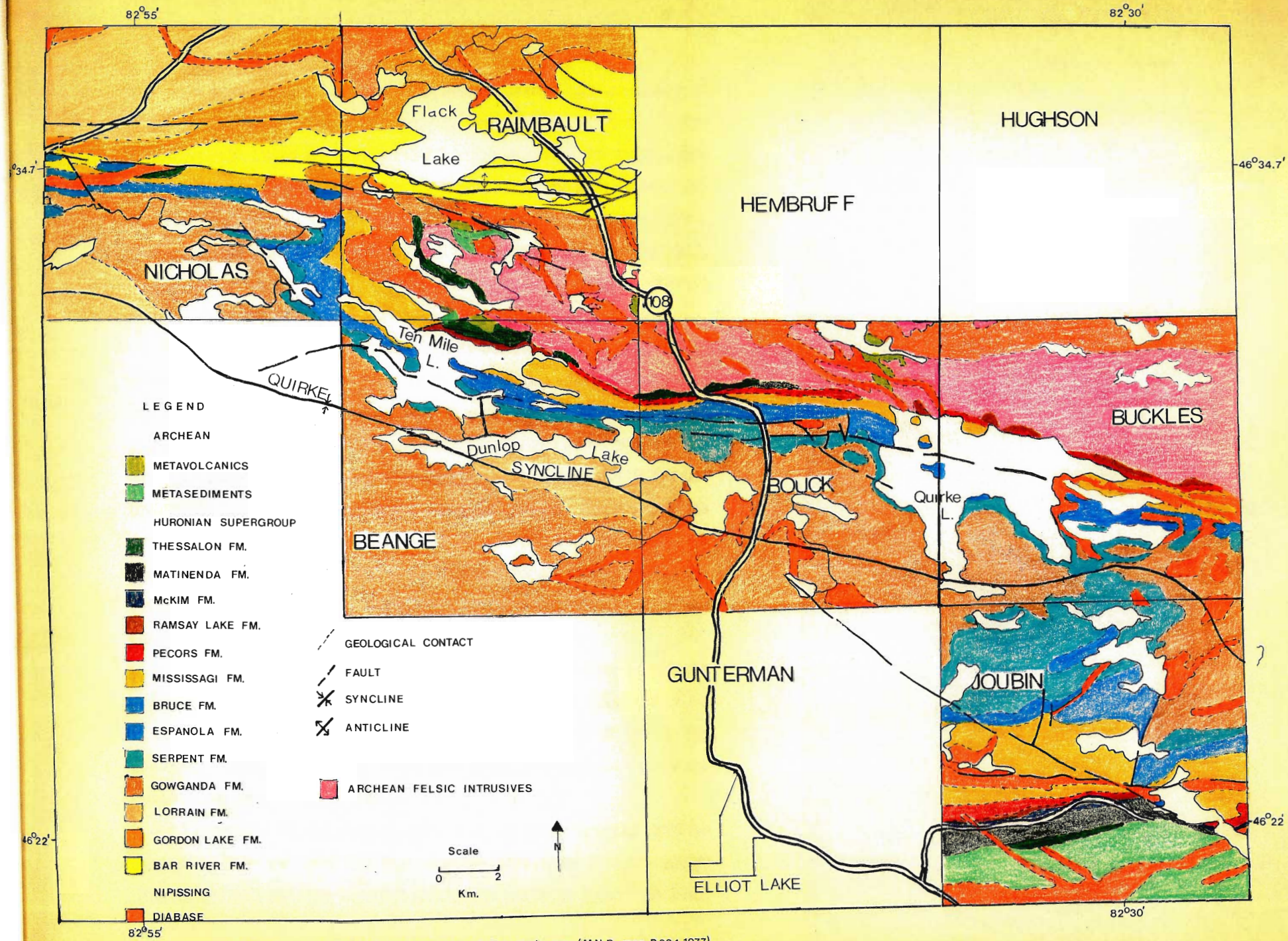

Figure 2 General geology map of the study area (M.N.R. map P.304,1977).

PREVIOUS WORK

The north shore of Lake Huron has been an area of intense geological activity since the discovery of copper at Bruce Mines in 1846. In the following year, Logan and Murray (1847) began to define the stratigraphy and structure of the strata between Sault Ste. Marie and Blind River. The succession, comprised predominantly of sandstones, conglomerates, slates, limestones and greenstones, was designated the Huronian (Robertson 1967; 1972). All of these rocks lie unconformably on older Archean granitic rocks of the Superior Province.

A second phase of exploration activity occurred between 1908-1925 when small sulphide and low-grade iron-formation deposits were exploited in the Blind River region. Intensified geological activity then began when Collins (1925) attempted to correlate the mineralized areas of the Bruce Mines with those of the Sudbury area. During this time, Collins (1925) sub-divided the Huronian from the base upwards as follows: The Bruce series consisting of the basal Mississagi Quartzite, the Bruce Conglomerate, the Bruce Limestone, the Espanola Greywacke, the Espanola Limestone and the Serpent Quartzite; the Cobalt Group including the Gowganda Formation and the Lorrain Quartzite (Table 1).

Recent geological investigations recognized the presence of uranium ore deposits at Elliot Lake. Following this discovery in 1953, extensive field work was undertaken by the Ontario Department of Mines, the Geological Survey of Canada and numerous mining companies to explore the Elliot Lake area for uranium ore.

Township mapping demonstrated that no post-Huronian granites occurred within the area and that uranium ores were probably placer deposits derived from pre-Huronian granites to the north (Robertson 1963;

Table 1 | COMPARISON OF STRATIGRAPHIC NOMENCLATURE USED FOR NORTH SHORE OF LAKE HURON MODIFIED FROM J. A. ROBERTSON (1967).

Logan and Murray as in Logan 1863	Coleman 1914	Collins 1925		Abraham 1953; 1956; and 1957.		Roscoe 1957 as in Pienaar 1963; and Roscoe 1960.		J. A. Robertson 1965a North of Murray Fault.		J. A. Robertson 1965a South of Murray Fault		Coleman 1914 and Collins 1925 South of Murray Fault		This report J. A. Robertson et al. 1968; 1969 a, b			Rock Types
Greenstone intrusions	Intrusions	Keweenawan	Olivine diabase	Keweenawan	Olivine diabase	Keweenawan	Olivine diabase	Keweenawan		Keweenawan		Keweenawan	Olivine diabase	Late PE	Late Mafic Intrusions		Olivine diabase
			Granite		Granite		Granite			Cutler Granite			Granite		Lamprophyre		Lamprophyre
			Diabase including Thessalon Greenstone		Diabase		Diabase	Nipissing		Nipissing			Diabase		Croker Island Complex		Diorite, granite, gabbro
															Cutler Granite		Granite, pegmatite
															Nipissing		Gabbro, diabase, granophyre
Upper formations	Upper Huronian	Cobalt Series	Upper formations	Cobalt Series	Lorrain	Cobalt Group	Upper formations	Cobalt Group	Upper formations	Cobalt Group	Not present in area mapped	Cobalt Series	Upper formations	Middle Precambrian (Middle PE)	Huronian Supergroup	Cobalt Group	Gordon Lake	Quartzite, siltstone
Upper slate conglomerate			Gowganda		Gowganda		Gowganda		Gowganda				Gowganda				Lorrain	Quartzite, arkose
																	Gowganda	Paraconglomerate, greywacke, quartzite
Limestone	Lower Huronian	Bruce Series	Serpent	Bruce Series	Serpent	Quirk Group	Serpent	Bruce Group	Serpent	Bruce Group	Serpent	Bruce Series	Serpent			Quirk Lake Group	Serpent	Quartzite
			Espanola		Espanola		Espanola		Espanola		Espanola		Espanola				Espanola	Calcareous and dolomitic siltstone, quartzite
Lower slate conglomerate			Bruce		Bruce		Bruce		Bruce		Bruce		Bruce				Bruce	Conglomerate
White quartzite			Mississagi		Upper Mississagi	Hough Group	Mississagi		Upper Mississagi		Upper Mississagi		Mississagi			Hough Lake Group	Mississagi	Quartzite
					Middle Mississagi		Pecors		Middle Mississagi		Middle Mississagi						Pecors	Argillite ± quartzite → schist
							Whiskey			Sprague Group			Ramsay Lake				Ramsay Lake	Conglomerate → schist
Chloritic slates						Elliot Group	Nordic		Lower Mississagi		Lower Mississagi	Sudbury Series	McKim			Elliot Lake Group	McKim	Argillite ± quartzite → schist
					Lower Mississagi		Matinenda (Upper)		Lower Mississagi		Pater Volcanics		Wanapitei				Matinenda	Quartzite, arkose, oligomictic conglomerate
Grey quartzite							Matinenda (Lower)										Salmay Lake	Mafic volcanic rocks, arkose
																	Gabbro Anorthosite	Gabbro Anorthosite
Laurentian Granite	Laurentian Granite	Algoman Granite		Algoman Granite		Algoman Granite		Algoman Granite		?		Schist Complex (Keewatin)		Early PE	Mafic Intrusions		Diabase
															Algoman Granite		Granite, etc.
	Sudbury Series	Sudbury Series		Keewatin		Greenstone		Keewatin							Keewatin – type metavolcanics and metasediments		Sedimentary rocks
	Granites and Keewatin	Schist Complex															Volcanic rocks

Frarey and Roscoe 1970). Roscoe (1957) presented data suggesting that the Espanola-Spanish sequence of the Quirke Group can be correlated with the Lower Huronian rocks at Elliot Lake; he subsequently developed a new nomenclature for the Blind River - Elliot Lake rocks (Table 1).

Major contradictions in stratigraphy and correlation resulted due to the amount of work by geologists on the north shore of Lake Huron. The Geological Survey of Canada and the Ontario Division of Mines developed their own nomenclature based on Collin's results, which in turn was revised by Roscoe (1957). Thus, in 1966, a committee was established to investigate the problems of Huronian stratigraphy and correlation, and to recommend a stratigraphic nomenclature. The names adopted by the Federal-Provincial Committee on Huronian Stratigraphy were based on the results of Collins (1925) and Robertson (1967) are given in Table 1.

Previous work has led to the view that the Elliot Lake volcanics were Keewatin in age. This assumption was based on the observation that Archean granites intruded the volcanics. Hence it was assumed that the volcanics were of the Keewatin era. Results published by Fairbairn *et al* (1969) have established that the Huronian was emplaced between 2500 and 2150 Ma ago. But radiometric dating of diabase bodies cutting the Huronian established their emplacement at 2155 ± 80 Ma (Van Schmus 1965); this age was confirmed by Fairbairn *et al* (1969). Furthermore, the granitic rocks of the basement were emplaced 2500 Ma ago (Fairbairn *et al* 1969) suggesting that these volcanic rocks were Keewatin in age. Rb/Sr dating performed on the Huronian basal volcanics yielded an isochron age of approximately 2005 ± 125 Ma, thus disproving the Keewatin age (Fairbairn *et al* 1969). However, this age indicates that Huronian volcanic rocks are younger than the Nipissing diabase. Field evidence illustrates that diabase

bodies cut Huronian rocks, thus volcanics must have been emplaced between 2500 and 2150 Ma ago. An isochron age of 2288 ± 87 Ma for the Gowganda Formation suggests that the Huronian volcanic rocks were emplaced between 2500 and 2288 Ma (Fairbairn *et al.* 1969).

GENERAL GEOLOGY

The Blind River - Elliot Lake region is a small part of the Southern Province of the Canadian Shield, which is composed of supracrustal rocks in a belt up to 443 km wide extending east from Minnesota to the Cobalt-Noranda region some 1288 km to the east. Boundaries for the Southern Province are defined by the Archean rocks of the Superior Province, younger Paleozoic units to the south, and by rocks of the Grenville Province to the east.

Rocks of the Huronian Supergroup occur in a folded belt occupying a 322 km by 64 km area along the north shore of Lake Huron where they attain a thickness of 12200 m. Maximum thickness of the Huronian rocks in the area is reached between Sudbury and Lake Huron, and they generally thin towards the north.

The Huronian Supergroup is composed of four major groups: Elliot Lake, Hough Lake, Quirke Lake and Cobalt (Fig. 2). Lithologically, most of these are sedimentary rocks, and these are punctuated by small exposures of volcanics which represent the lowest member of the Elliot Lake Group. Rocks of the Elliot Lake Group comprise the lowermost unit of Huronian stratigraphy and lie with unconformity upon the Archean granitic basement. This basal member consists of several thousand metres of interbedded sandstone and mafic extrusive and related intrusive rocks. Volcanic rocks are generally massive in nature but do show amygdaloidal, schistose and

porphyritic textures. These rocks constitute the lowermost unit of the Huronian Supergroup.

Intercalated with the volcanic unit is the Matinenda Formation consisting of feldspathic sandstones and conglomerates. The sandstone unit is predominantly composed of quartz and feldspar grains in a matrix of micas, chlorite and quartz. Locally, lenses of oligomictic quartz-pebble conglomerate occurred with a coarse grained matrix of quartz plus feldspar and micaceous minerals. It is this conglomerate unit that contains the ore bearing uranium-thorium minerals that established Elliot Lake as a major uranium producer. Mineralogically and texturally, the sandstones are immature to submature and grain size plus sorting increases upwards in the formation, indicating a turbulent environment of deposition for this unit (Card and Palonen 1976)(Table 2).

Lying conformably above the Matinenda Formation with a gradational contact is the McKim Formation, which consists of argillites, siltstones, greywackes and metapelites. The pelites have various proportions of micas, chlorites, fine grained quartz and feldspar grains, along with accessory sphene and opaque minerals. Bedding thickness in the McKim Formation averages 5 cm. Sedimentary structures present include laminated and graded bedding yielding top data and current directions. In addition, cross laminations suggest deposition by turbidity currents (Card and Palonen 1976).

Conformable with the underlying McKim Formation is the first unit of the Hough Lake Group; the Ramsay Lake Formation. This formation is predominantly a paraconglomerate and can be divided into two parts: the lower half consists of 15-20% clasts and boulders in a coarse-grained greywacke matrix. The upper half consists of 5-10% clasts in a

TABLE 2. Lithologies and Depositional Environments of Rocks of the Huronian Supergroup.

GROUP	FORMATION	LITHOLOGY	DEPOSITIONAL ENVIRONMENT
Elliot Lake	Pecors Lake, Dollyberry Lake, Crazy Lake		
	Matinenda	Sandstone, Conglomerate	Fluvial
	McKim	Argillite, Siltstone, Greywacke, Metapelite	Turbidity Current
Hough Lake	Ramsay Lake	Conglomerate	Debris Flow, Glacial
	Pecors	Argillite, Siltstone, Sandstone	Turbidity Current
	Mississagi	Arenite, Siltstone, Sandstone	Fluvial
Quirke Lake	Bruce	Conglomerate	Debris Flow, Glacial
	Espanola	Limestone, Dolostone, Siltstone	Shallow Marine, Turbidity Current
	Serpent	Quartzite, Siltstone, Argillite, Conglomerate	Shallow Water
Cobalt	Gowganda	Conglomerate, Greywacke, Argillite, Siltstone	Glacial
	Lorrain	Conglomerate, Quartzite, Siltstone, Orthoquartzite	Fluvial, Beach
	Gordon Lake	Siltstone, Quartzite	Shallow Water
	Bar River	Orthoquartzite	Beach

subgreywacke matrix. A debris flow origin is normally interpreted to account for the Ramsay Lake Formation (Card and Palonen 1976). However, Long (1981) suggested this formation may represent glacial outwash deposits, and the greywacke-subgreywacke micticte sequence may be of glaciogenic origin, and the occurrence of scattered pebble horizons in the overlying unit represent ice rafted material (Table 2).

Higher in the Huronian succession, the next unit is the Pecors Formation, which is conformable with the subjacent formation. This formation is composed of argillites, siltstones and sandstones. The lower 30.5 m is comprised of thin-bedded, laminated argillite and siltstone. Above this, are approximately 30.5 m of interbedded siltstone, greywacke and feldspathic protoquartzite. Planar crossbeds are visible in the upper 30.5 m, while in the lower part ripples, cross laminations and ball-and-pillow structures are visible, which suggests deposition by turbidity currents (Card and Palonen 1976).

The youngest member of the Hough Lake Group, the Mississagi Formation, exhibits planar bedding and cross stratification within medium to coarse grained feldspathic arenites. Interbedded in these units are siltstones and pebbly sandstones with scarce quartz and granite clasts. No systematic cyclicity was noted in the sedimentary rock types in the Missisagi as was reported by Palonen (1973), and upward fining cycles are lacking in the formation (Pettijohn 1970; Long 1977). The bulk of the Mississagi Formation was deposited in a fluvial environment; principally from bed load and mixed load streams. These rivers were marked by braided stream patterns which were characterized by high width to depth ratios (Long 1977).

Succeeding the Mississagi Formation is the youngest unit of the Quirke Lake Group, the Bruce Formation, which is dominated by a polymictic conglomerate consisting of subangular cobbles, boulders and clasts of granite, gneiss and diabase set in a greywacke matrix. It appears that this lithology is similar to that of the Ramsay Lake Formation, thus the same mode of deposition may apply to this unit (Card and Palonen 1976), although this conglomerate unit may be assigned a glacial origin (Long 1977; Young 1973).

Conformably overlying the Bruce Formation, is the Espanola Formation, composed of limestone, dolostone and siltstone. The limestone and dolostone units represent shallow marine deposits (Young 1973), whereas the intervening siltstone units containing ripple marks and cross and graded bedding, indicate deposition in deep waters from turbidity currents (Young 1973).

The next unit of the Quirke Lake Group is the Serpent Formation which consists of feldspathic quartzite, siltstone, argillite and conglomerate. Predominantly this unit is a well bedded quartzite to feldspathic quartzite which weathers white. This unit was formed in a shallow water environment, and was subsequently subjected to low grade metamorphic conditions causing recrystallization of the grains and the formation of the massive quartzite (Card and Palonen 1976).

The youngest member of the Huronian Supergroup is the Cobalt Group, resting unconformably on the underlying Quirke Lake Group. The Gowganda Formation is 1,000-1,500 m thick and is composed of massive conglomerate, greywacke, arkose and siltstone. Paraconglomerate in the lower half of the formation comprises clasts, cobbles and boulders of granitic rocks, diabase, greenstone and quartz in a greywacke matrix. The

textural immaturity indicates that deposition occurred during glacial conditions (Symons 1975; Card and Palonen 1976).

Succeeding the Gowganda Formation, is the Lorrain Formation. This member consists of conglomeratic, feldspathic quartzites, arkoses with interbedded siltstones, quartzites and orthoquartzites. The conglomerate of the Lorrain Formation has a fluvial origin, whereas the upper quartzite units represent beach deposits (Long 1981).

Conformably overlying the Lorrain Formation is the Gordon Lake Formation, consisting of 305 m of siltstone and argillaceous quartzite. These lithologies and the sedimentary structures which include current ripples, cross-bedding and desiccation cracks point to a shallow water environment (Card and Palonen 1976).

Atop the Cobalt Group, is the youngest unit; the Bar River Formation. This lithological unit is characterized by a massive to well-bedded orthoquartzite which is indicative of beach deposits (Long 1981).

REGIONAL GEOLOGY

The volcanics, in this study, belong to the Thessalon Formation and comprise the Pecors Lake, Dollyberry Lake and Crazy Lake volcanic belts (Robertson 1971). Within Joubin Township between latitudes 46° 21.5'N and 46° 22' N and longitudes 82° 30'W and 82° 7' W are the Pecors Lake volcanics (Figs. 2, 3). This belt is overlain by sediments of the Matinenda Formation, and is bounded to the east, west and south by Archean metasediments, and to the south by Nipissing intrusives. Exposures range in area from small moss covered showings to large cliff-like structures with differences in relief up to about 30.5 m.

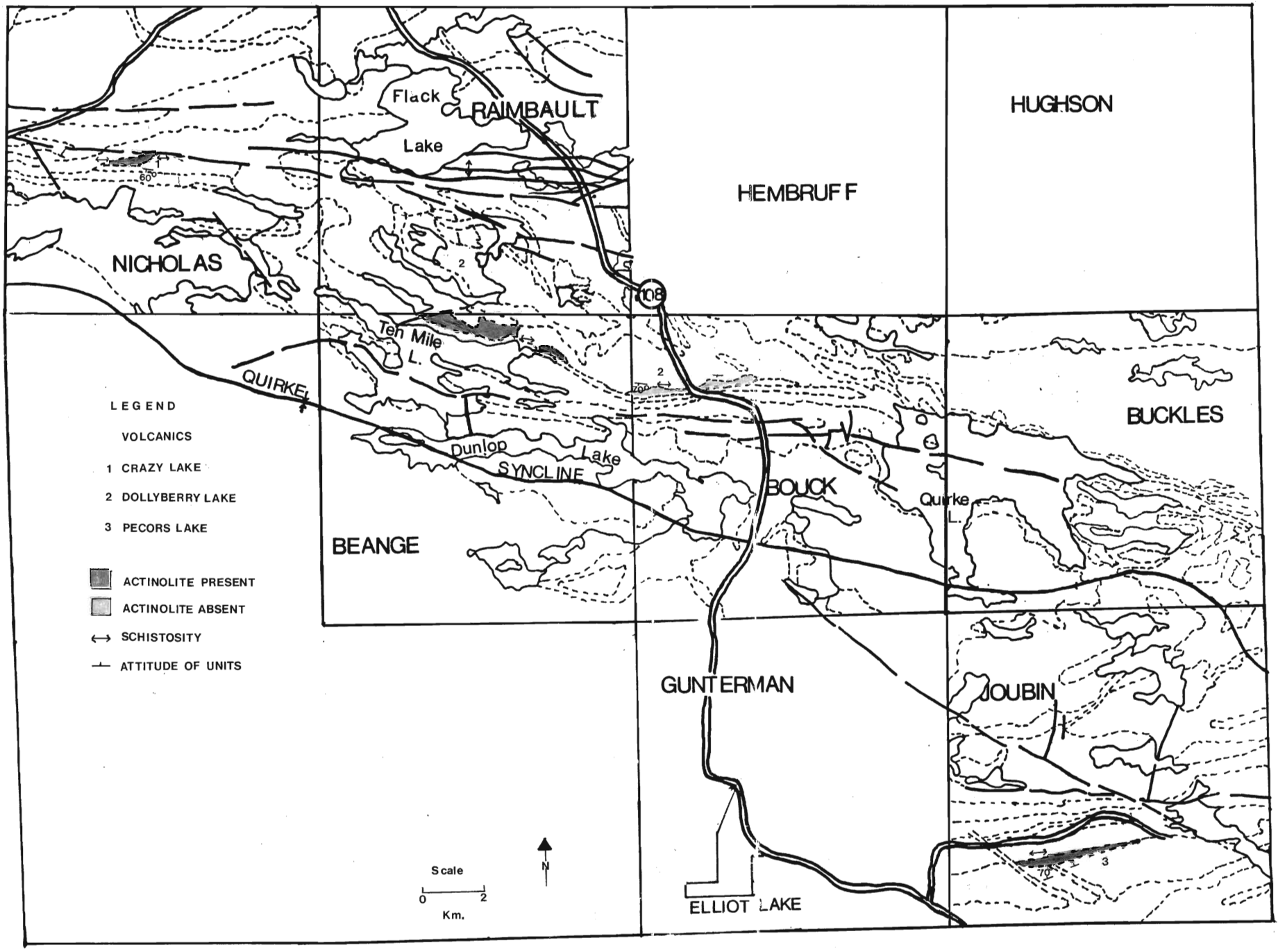

Figure 3 Location of the different Elliot Lake volcanic belts (M.N.R. map P.304, 1977)

Mafic volcanics are grey-green to dark green on fresh surfaces; the rocks weather grey to brownish grey. These fine-grained rocks are generally massive in structure but commonly contain amygdules and veins of plagioclase and quartz (Fig. 4). Plagioclase also occurs in small glomerophenocrysts and quartz exists in small stringers, amygdules and infill large vugs.

Lava piles trend N70° E to N120° E and dip to the north. Schistosity is rare, but was observed in the western part of the belt trending approximately east-west. Small-scale brecciation occurs in angular fragments of volcanic material, which is incorporated in plagioclase-orthoclase veins (Fig. 5). Amygdules are subspherical structures that are microscopic to a few centimeters in size and are infilled with quartz, calcite, albite, minor biotite, but the most common infilling is chlorite.

Glomerophenocrysts of plagioclase exist in close association with chlorite amygdules (Fig. 6). Bennett (1982) reported that amygdules are widely dispersed in the centre of flows. However, individual flows were difficult to distinguish because no substantial flow textures were visible. It is probable that amygdules are the result of degassing. Shallow igneous intrusions in close proximity indicate that the rocks degassed in a low pressure environment. Vugs support this theory. Thus, amygdules are likely to be microlitic degassing structures (Fig. 7).

Samples of rocks representing the Dollyberry Lake volcanics were obtained from outcrops situated between latitudes 46° 30.5' and 46° 32' and longitudes 82° 38.5' and 82° 47.5'. These volcanics trend east-west with bedding steeply dipping south. Lying upon argillites of the Pecors Formation are conglomerates of the Mississagi Formation. Dollyberry Lake volcanics are bound to the north by Archean and Nipissing intrusives. To the

Figure. 4. Small outcrop illustrating quartz veins and stringers (1 m=50 cm).

Figure 5. Small-scale brecciation as volcanic material is incorporated into plagioclase-orthoclase veins (light-coloured material).

Figure. 6. Glomerophenocrysts of plagioclase in close association with chlorite amygdules (1 m=40 cm).

Figure 7. Vugs occurring in close association with chlorite amygdules.

east, the volcanics are truncated by part of the Quirke Lake Fault system and to the west by sediments of the Matinenda Formation.

The outcrops extend for approximately 15 km along the continuous ridge that rises to 473 m in Bouck Township. The rocks are dark grey-green to black and are massive to amygdaloidal with local porphyritic textures and irregular showings of pyrite. Subspherical amygdules are filled primarily with chlorite, but quartz commonly substitutes for chlorite near the eastern end of the belt. Plagioclase occurs in porphyritic flows and veins up to 1.22 m thick. Schistosity is also more pronounced in these rocks which trend east-west, parallel to the Quirke Lake Fault.

A common component throughout the volcanic units of this belt are plagioclase glomerophenocrysts; lumps of crystals arranged in random orientation and a seriate texture, with phenocrysts up to 4mm in length and 1.5mm in width (Fig. 8). Sixty meters west of Highway 108, an east-west trending contact between granites and volcanics was found. This contact was traced to the west over a distance of 46 m. Plagioclase glomerophenocrysts are closely associated with amygdules, while plagioclase veins are not common features.

Dollyberry Lake volcanics continue west and are visible on the north shore of Ten Mile Lake overlying sediments of the Mississagi Formation (Figs. 2, 3). These volcanic rocks are grey-green to dark green and exhibit the same textures. Chlorite amygdules are up to 8mm in length and 2mm in width. Plagioclase glomerophenocrysts have similar characteristics, and schistosity maintains its east-west trend. Pyrite is rare and plagioclase veins are not dominant. This volcanic belt extends for 3.5 km and is cut-off to the east and northeast by the Quirke Lake Fault. North of this fault situated between Hyphen and Callinan Lakes, metasediments were observed,

rather than Dollyberry Lake volcanics, as indicated by Ontario Department of Mines map 2015 and Ministry of Natural Resources Preliminary map 304.

A small sliver of Dollyberry Lake volcanic rocks is situated at the northeast end of Ten Mile Lake (Figs. 2, 3). This section is approximately 0.5 km by 0.2 km and elevation attains 473 m. Outcrops reveal the same characteristics as the large belt on the north shore of Ten Mile Lake, but vugs are a dominant texture, present in both massive and schistose samples (Fig. 9).

Within Nicholas Township, situated between latitudes 46° 34.4' and 46° 34.7' N and longitudes of 82° 53.8' and 82° 55.0' W, is the Crazy Lake volcanic belt (Figs. 2, 3). This region is less extensive than the area indicated by the Ontario Division of Mines map 2347. Crazy Lake volcanics are grey-green to dark green, fine-grained, massive rocks containing chlorite amygdules and minor quartz veins. This formation is truncated to the north by the Flack Lake Fault and bounded to the west by Nipissing intrusives.

Schistosity is a rare feature in outcrops and topographic relief ranges from 458 to 473 km. The Crazy Lake belt is along strike from the Dollyberry Lake volcanics and bedding dips steep to the south.

STRUCTURE

The Quirke Lake Syncline represents the major structural feature in the study area. This fold is a result of tectonic activity which occurred after the deposition of the Cobalt Group (Robertson 1961), but prior to intrusion of Nipissing rocks (Robertson 1961, 1963). The Quirke Lake Syncline trends north 70-85 degrees west, plunges west approximately 1 degree, and extends a distance of 55 km from just west of Nicholas

Figure. 8. Plagioclase glomerophenocrysts arranged in random orientation and depict a seriate texture.

Figure 9. Volcanics show vugs and an east-west trending schistosity.

Township to 10 km east of Joubin Township. The north limb of the fold dips south 15-60 degrees, while the south limb dips north 20-30 degrees. Hence, complementary anticlines exist north and south of this shallow syncline, creating an asymmetric fold which has its axial plane lying closer to the north limb and its axial surface oriented east-west.

Closely related to the Quirke Lake Syncline are the Flack Lake, Quirke Lake, and Pecors Lake Faults. These faults are contemporaneous with folding (Robertson 1961, 1963; Card *et al.* 1972) and occurred during the Penokean Orogeny prior to the intrusion of the Nipissing diabase (Card et al. 1972).

In Beange and Bouck Townships, the Quirke Lake Fault strikes approximately east-west. This fault is a thrust on the north limb of the Quirke Syncline with the direction of slip to the north, and vertical displacement ranges from 28-76 m (Pienaar 1963). Similarly, the Flack Lake Fault trends east-west causing north-south shortening, and vertical displacement up to 1220 m has been proposed (Robertson 1971).

To the west, the Dollyberry Lake volcanics are truncated by a northwest trending segment of the Quirke Lake Fault. This structural feature represents a wrench fault, one which has caused sinistral displacement of volcanics to the north (Robertson 1971).

In Joubin Township, to the west of the Pecors Lake volcanics, is the Pecors Lake Fault, a wrench fault with lineament striking N60° W, parallel with the majority of diabase intrusions. This fault represents dextral movement, as rocks west of the fault are displaced relative to rocks east of the fault (Robertson 1971, 1977).

PETROGRAPHY

PECORS LAKE FORMATION

The metavolcanics of the Pecors Lake Formation have undergone ubiquitous alteration to secondary mineral assemblages. Chlorite is the predominant secondary mineral, whereas actinolite and biotite pervade all rock samples. Actinolite exhibits pale yellow to pale green pleochroism in single prismatic grains, prismatic aggregrates, single acicular fibers or aggregates of fibers. Chlorite occurs as single prismatic grains or prismatic aggregates and is pleochroic pale yellow - green to green, and displays predominantly blue grey to dark green birefringence, while some crystals exhibit anomalous purple interference colours.

All rocks of the Pecors Lake Formation contain albitized plagioclase microlites of seriate form. Some prismatic crystals have been completely converted to chlorite, and to a lesser extent calcite. Atoll textures are evident to varying degrees in all samples as small patchy aggregates of chlorite and minor actinolite have formed within albitized plagioclase laths (Plate 1). Some laths also contain small amounts of epidote and in some instances complete epidotization of plagioclase has occurred (Plate 2).

The groundmass enclosing microlites is an aphanitic granular association of chlorite, actinolite, biotite, epidote, quartz, albite, titanomagnetite and sphene. Accessory minerals include calcite, hematite and clinozoisite (Plate 3). Biotite occurs at the edges of chlorite crystals and grows inwards to form single prismatic grains or prismatic aggregates (Plate 4).

A few samples contain amygdules, in which chlorite is ubiquitous, occurring with calcite, sphene, hematite and specular hematite to surround

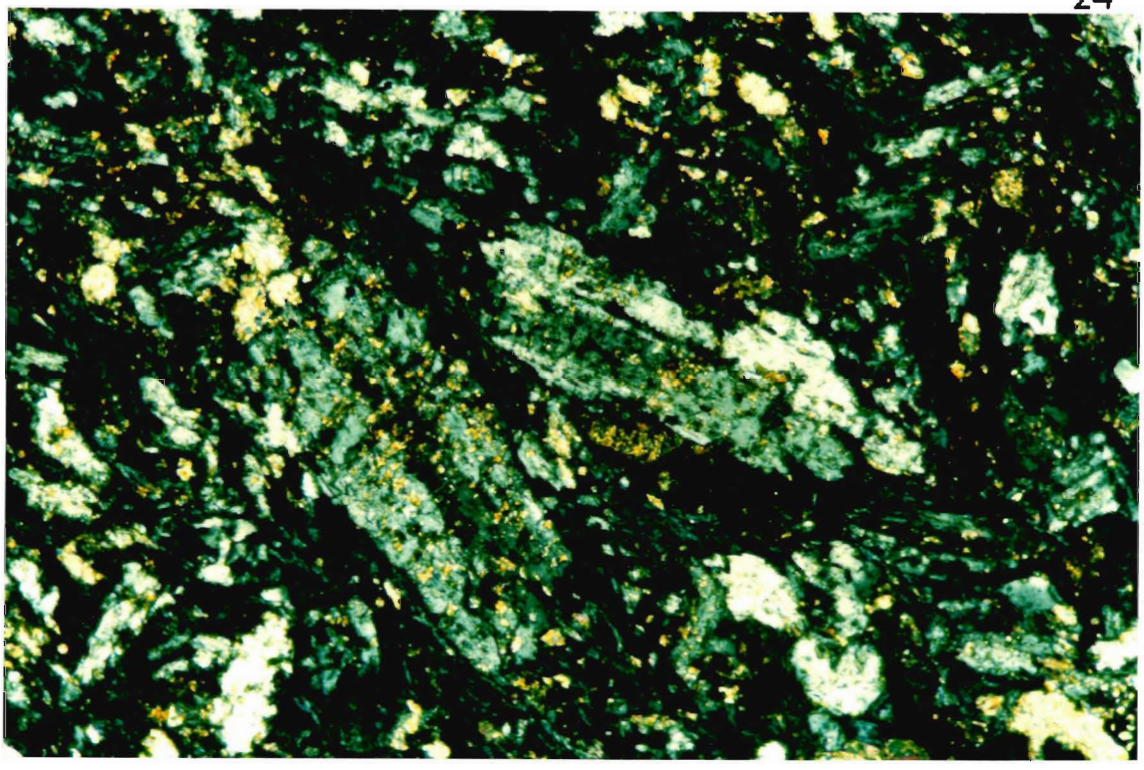

Plate 1. Chlorite and minor actinolite atoll-texture forming in plagioclase microlites (3.2 cm=0.5 mm).

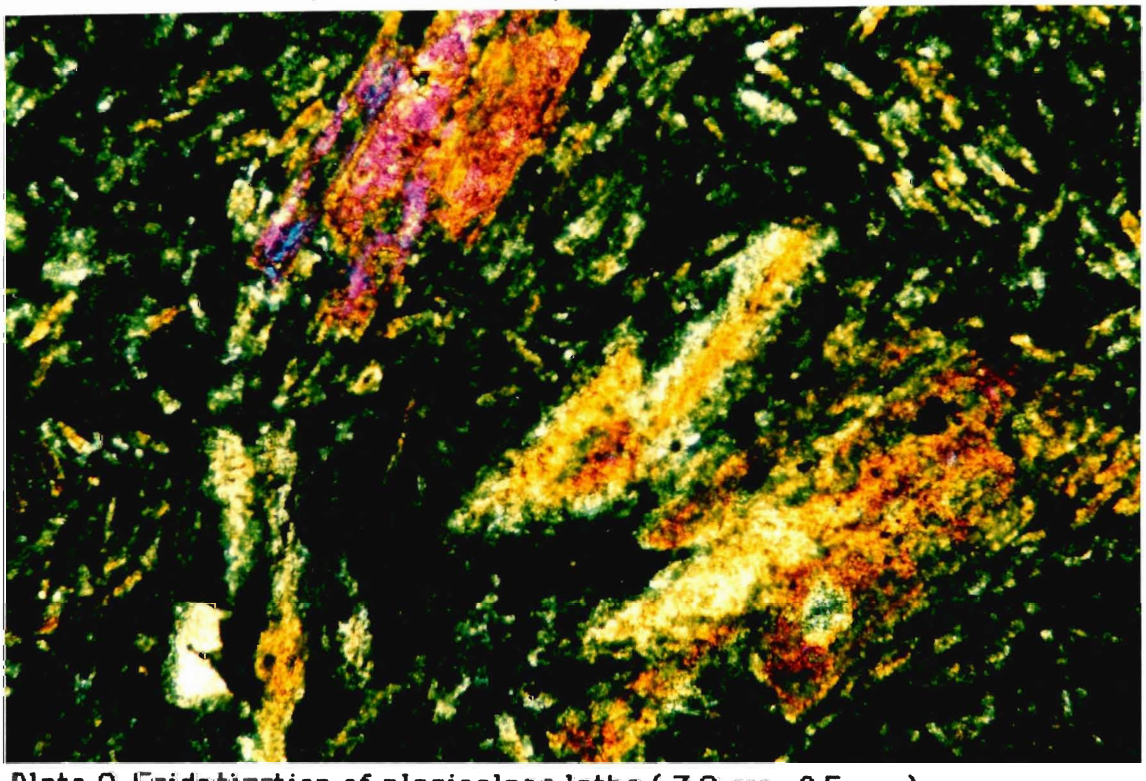

Plate 2. Epidotization of plagioclase laths (3.2 cm = 0.5 mm).

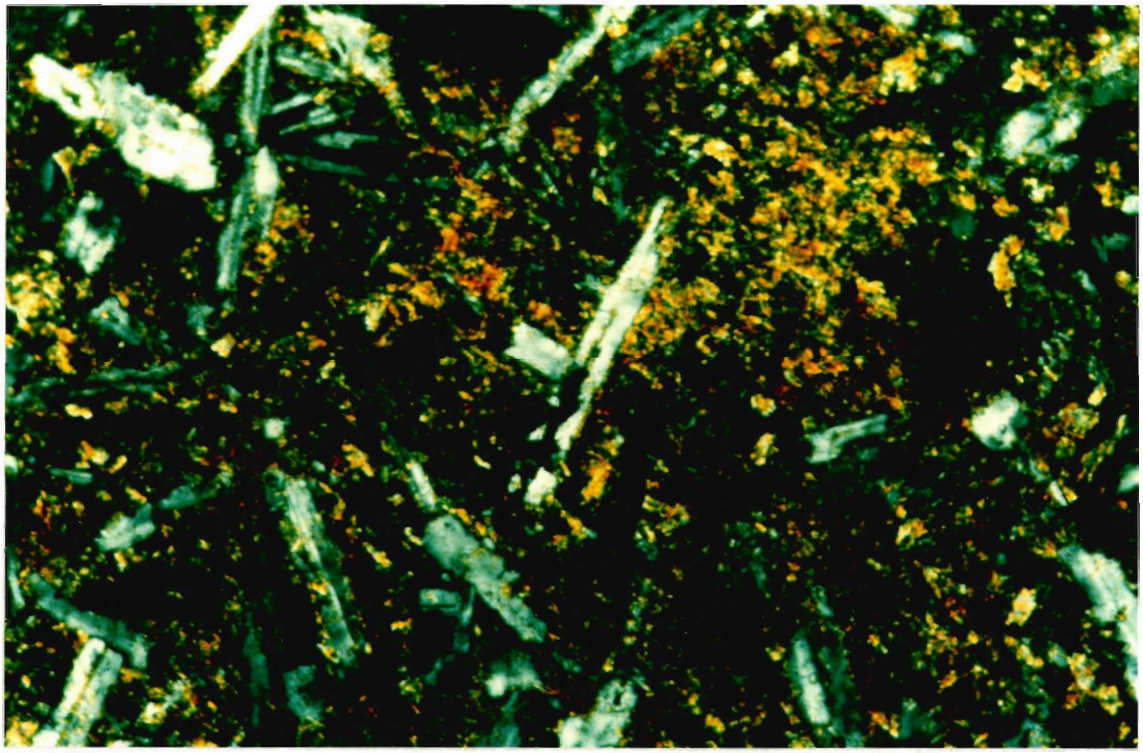

Plate 3. Interstitial arrangement of metamorphic greenschist-mineral assemblage (chlorite, actinolite, biotite, epidote, quartz, albite, sphene, calcite, hematite)(3.2 cm=0.5 mm).

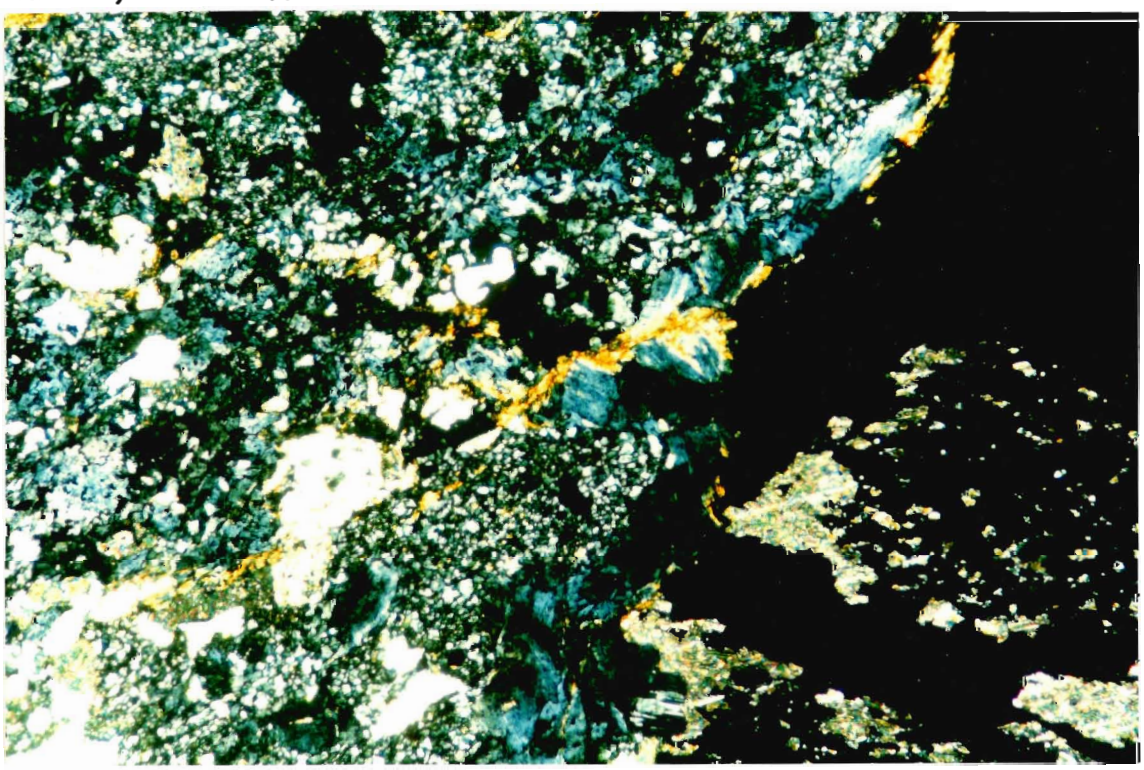

Plate 4. Peripheral biotite replacement of chlorite (3.2 cm= 0.5 mm).

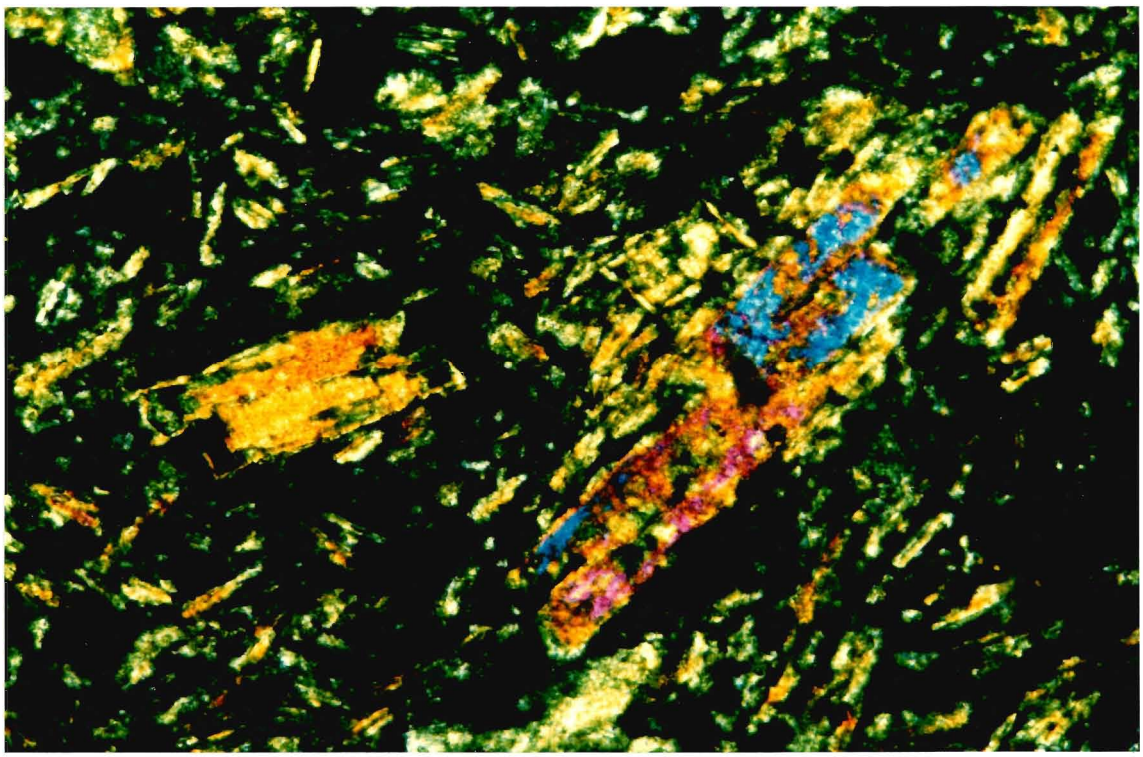

Plate 5. Albitization and subsequent epidotization of plagioclase laths (3.2 cm=0.5 mm).

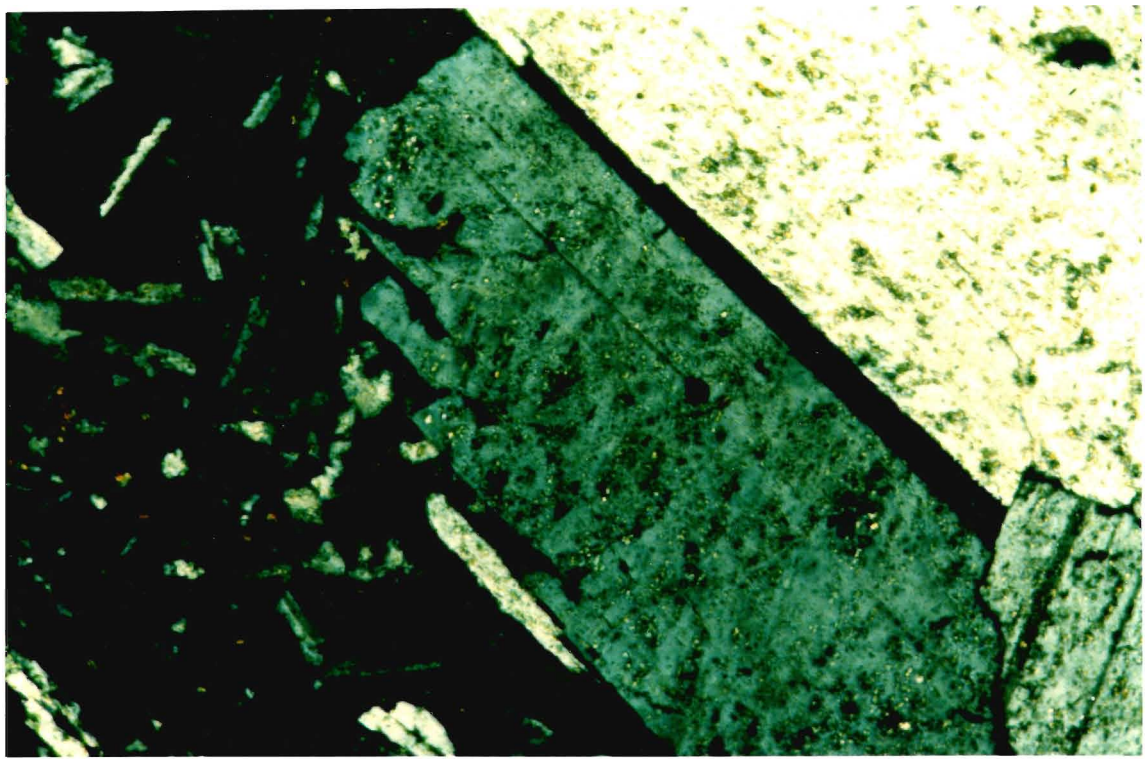

Plate 6. Plagioclase glomerophenocrysts depicting typical atoll texture (3.2 cm= 0.5 mm).

represents a large fraction of the modal percent in Dollyberry Lake rocks ranging from 15 to 65 percent, but averages 45 to 60 percent.

Amygdules up to 5 mm x 2 mm in width are chlorite filled with minor albite, biotite and quartz. Small porphyroblastic veins and vesicles contain quartz, albite and minor amounts of chlorite and biotite. Secondary hematite forms a poikiloblastic texture with chlorite, biotite and epidote, whereas acicular chlorite fibers are found within quartz crystals (Plate 7); in addition inclusions of hydrogarnet were observed within poikiloblastic chlorite residing in amygdules of many samples (Plate 8).

Situated on the north shore of Ten Mile Lake, west of longitude 46°, are the amphibole bearing Dollyberry Lake volcanics. Actinolite is pleochroic pale yellow to pale green and exhibits the prismatic and acicular crystal habits common in the Pecors Lake Volcanics. Seriatic albitized plagioclase microlites are evident in most samples and illustrate the progressive replacement of plagioclase by chlorite, epidote and actinolite. Complete conversion of plagioclase laths to chlorite and epidote has occurred in a few samples.

Interstitial to microlites and relict plagioclase laths are the same minerals evident in previous formations, but clinozoisite is more prevalent in these amphibole bearing volcanics than in samples from other formations. Only a few rocks show veinlets; normally containing a fine-grained assemblage including quartz, albite, epidote, clinozoisite, chlorite and calcite.

Amygdules are filled with a variety of minerals of which chlorite is dominant. It is associated with other greenschist-grade secondary phases and is berlin blue-green and exhibits undulatory extinction. Actinolite is visible rimming and cross-cutting chlorite (Plate 9). In one sample,

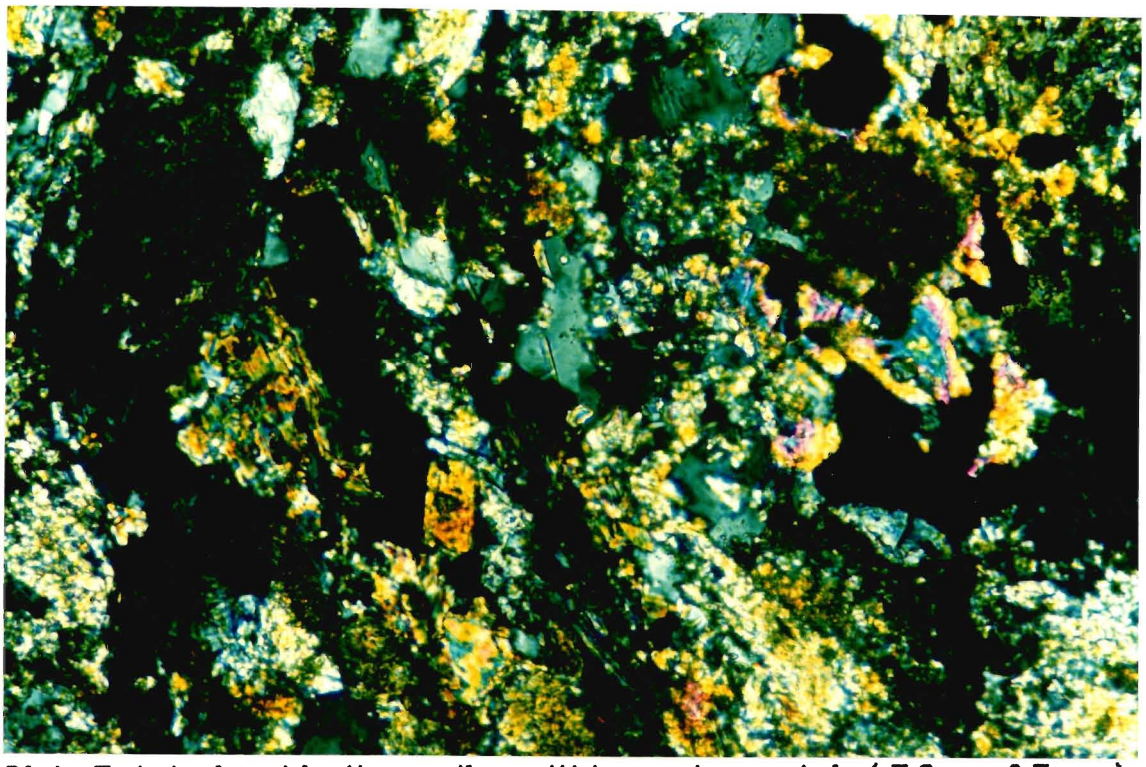

Plate 7. Acicular chlorite needles within quartz crystals (3.2 cm=0.5 mm).

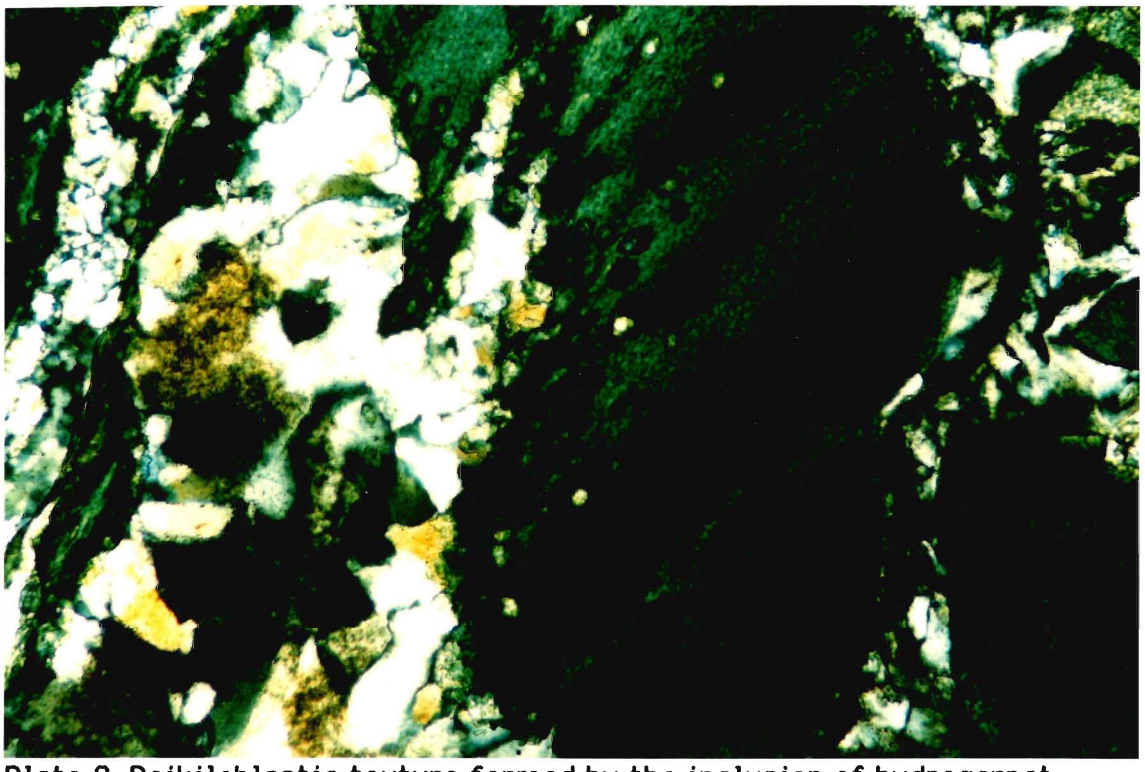

Plate 8. Poikiloblastic texture formed by the inclusion of hydrogarnet crystals within chlorite amygdules (3.2 cm= 0.5 mm).

amygdules were filled with only calcite, but this rock was collected just east of longitude 46° and did not carry actinolite.

CRAZY LAKE FORMATION

The Crazy Lake Formation represents the smallest belt of the Huronian volcanics studied, but it contains a wide variety of textures. Seriate plagioclase microlites are evident in many samples. Albitization has occurred and patches of chlorite are visible within the lath structures. Complete conversion of plagioclase has occurred in other samples as relict laths now contain mostly epidote with minor chlorite and albite. Interstitial to these microlites is an interlocking arrangement of chlorite, actinolite, biotite, epidote/clinozoisite, titanomagnetite, sphene ± hematite (Plate 10). Amygdules are filled with a variety of constituents. Chlorite is present in all amygdules, whereas actinolite either rims or crosscuts existing chlorite. Amygdules also contain a random arrangement of chlorite, actinolite and epidote, or consist only of chlorite. Some samples contain amygdules that were filled with epidote rimmed by chlorite and actinolite.

Rocks from the western end of the Crazy Lake belt have no plagioclase microlites. These samples exhibit an aphanitic, interlocking mosaic of chlorite, actinolite, biotite, epidote, clinozoisite, albite, quartz, sphene, titanomagnetite and hematite. A few samples exhibit an east-west lineation defined by the preferred orientation of chlorite, biotite and actinolite. Small veins of clinozoisite are situated between the chlorite, biotite and actinolite metadomains, and these rocks also contain small veins of fine-grained quartz.

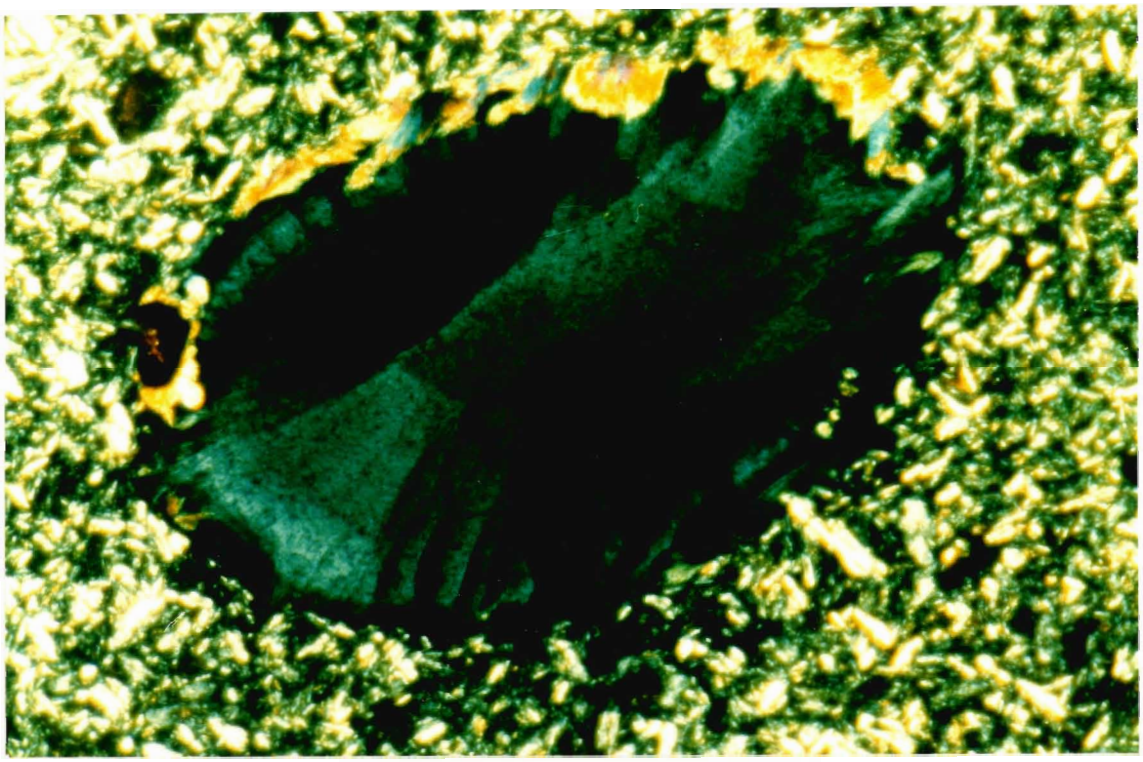

Plate 9. Amygdule showing actinolite rimming and incising chlorite (3.2 cm=0.5 mm).

Plate 10. Greenschist metamorphic-mineral assemblage showing sphene replacing opaques (3.2 cm= 0.5 mm).

METAMORPHISM

Volcanic units in the thesis area are representative of the greenschist facies. The lava piles of these regionally metamorphosed rocks suggest a temperature range of 325° C (Schiffman and Liou 1980) to 450° C (Studemeister 1983; Liou *et al.* 1974) and a pressure regime of 2-3 kilobars (Studemeister 1983) (Fig. 10).

Rocks of the Pecors Lake, Dollyberry Lake and Crazy Lake Formations have mineral assemblages (albite - actinolite - chlorite - epidote/clinozoisite) indicative of low grade metamorphism. Original mineralogy is mainly destroyed but some primary textures are evident.

These metamorphosed mafic rocks carry plagioclase with a consistent anorthite composition. Plagioclase in rocks of the greenschist facies is nearly pure albite (An < 7, Liou *et al.* 1974). Plagioclase laths not completely albitized have been partly replaced by chlorite, epidote, calcite or actinolite. In a few samples, plagioclase microlites have been totally epidotized or chloritized. Epidote is a common greenschist facies mineral forming mainly at the expense of plagioclase. Ca is liberated from plagioclase crystals and is a source for local enrichment that occurs in lavas (Smith 1968). Released Ca is then consumed to form calcium bearing minerals such as actinolite, epidote, sphene and calcite. Albitization of plagioclase occurs as Na^{+} ions from a reservoir in the groundmass enter the plagioclase framework and infill the vacancies left by departing Ca^{2+} ions. The breakdown of titanomagnetite releases Ti into the magma which combines with liberated Ca to form sphene. The formation of epidote and hematite suggest high fluid pressures of H_2O and O_2 and suppress Fe-enrichment (Osborn 1959, 1962; Jolly 1978) and ubiquitous biotite forms

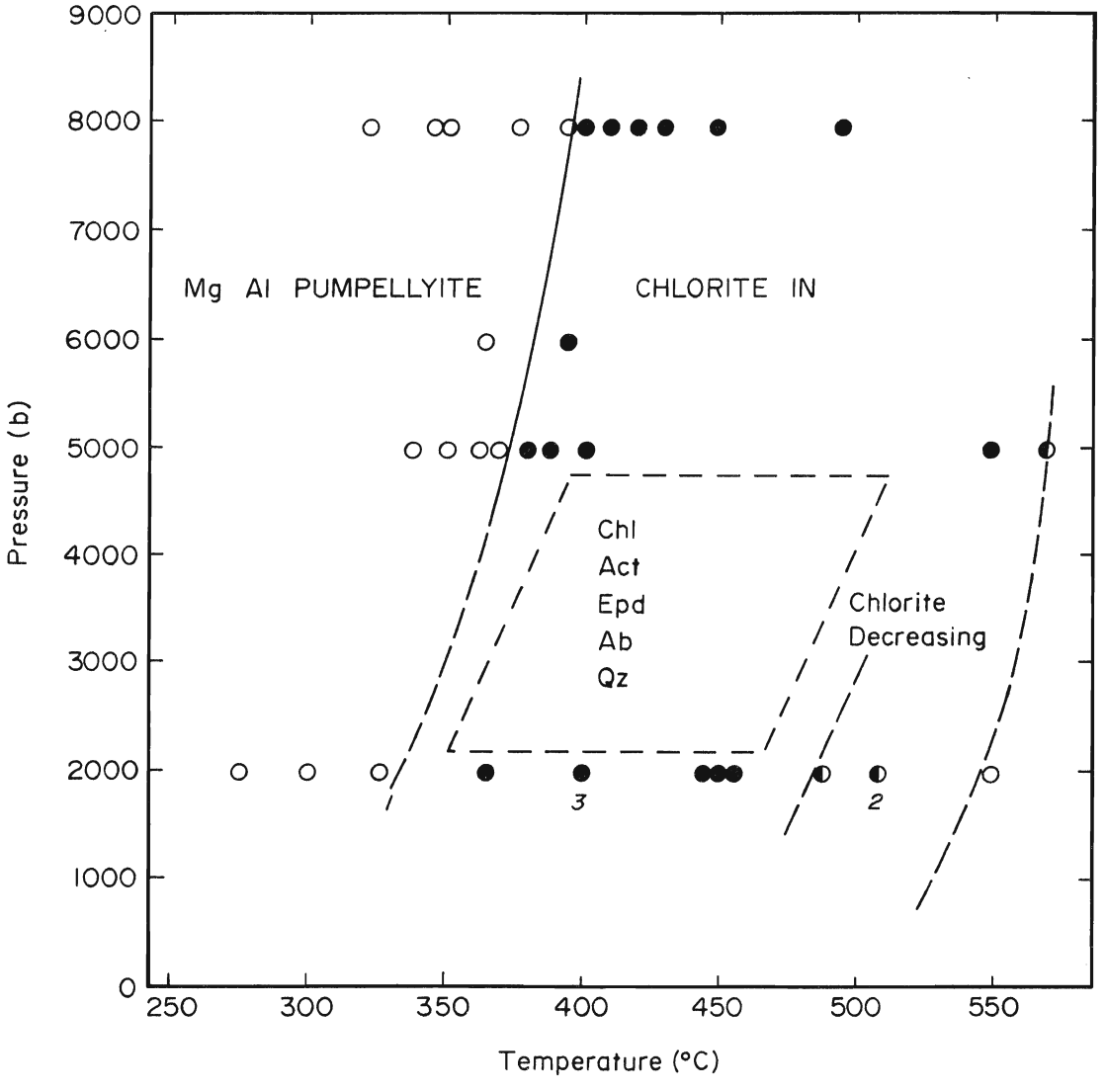

Figure 10. Temperature and pressure conditions of greenschist metamorphism acting on Elliot Lake volcanic rocks (1, Schiffman and Liou 1980; 2, Studemeister 1980 and Liou *et al.* 1974; 3, this study).

as a result of K_2O being introduced into the system by alkali intrusions (Jolly 1978).

Samples collected adjacent to intrusions illustrate the presence of clinozoisite. Clinozoisite is distinguished from epidote by its clear nature in plane light and anomalous blue interference colour.

The presence of calcite, epidote and chlorite suggests high pCO_2, conditions common in greenschist facies (Jolly 1981; Davison 1984). Conditions favouring the formation of sphene indicate lower pCO_2 conditions and restricts the amount of calcite formed (Schuiling and Vink 1967).

Volcanic rocks of the Dollyberry Lake Formation collected east of longitude 46° contain no actinolite. Thus, these greenschist facies rocks can be divided into two zones on the basis of the presence or absence of this mineral (Table 3).

Mineral assemblages of these rocks contain variable amounts of biotite, thus the metamorphic volcanics can be assigned to the biotite zone of the Greenschist Facies (Smith 1969; Winkler 1979).

The lower temperature constraint of 325° C is imposed due to the absence of pumpellyite (Schiffman and Liou 1980) and the presence of abundant chlorite with epidote, actinolite and albite suggests an upper limit of 450° C (Studemeister 1983; Liou *et al*. 1974) (Fig. 10). These volcanic rocks are covered by 6.7 (Meyn 1973) to 12.2 km (Frarey and Roscoe 1970) of Huronian sedimentary rocks. This depth corresponds to burial of approximately 2.4 to 4.7 kb (Miyashiro 1973). Thus, Elliot Lake volcanic rocks fit the constraints cited by Schiffman and Liou (1980), Studemeister (1983) and Liou *et al*. (1974) as evident in Figure 10.

Table 3. Metamorphic mineral assemblages found in the rocks of the study area.

Minerals	Zone 1	Zone 2
Actinolite	X	
Albite	X	
Chlorite	X	X
Epidote/Clinozoisite	X	X
Biotite	X	X
Calcite	X	X
Opaques	X	X
Sphene	X	X
Hematite	X	X

Note: X= present.

PETROCHEMISTRY

A total of 110 rocks were collected from the study area, of which 55 were analyzed for major and minor oxide contents. Samples that have undergone severe alteration, or which contained amygdules, vesicles or other secondary features were used for petrographic analysis. Furthermore, volcanic rocks which yield oxide totals lower than 97.50 and exceeding 102.50 were excluded from consideration. Forty rocks that were analyzed for major oxides were also selected for trace element determinations. The major oxides include SiO_2, Al_2O_3, FeO, MgO, CaO, Na_2O, K_2O, TiO_2, MnO and P_2O_5. Trace elements analyzed are Ba, Ce, Cu, Ni, Rb, S, Sr, Y, Zn and Zr.

Major and trace element analyses were performed by the author using a Philips 1450/20 sequential automatic X-ray fluorescence spectrometer. Homogeneous glass discs were used for major element oxide determination while pressed powder pellets were analyzed for trace element analysis. Major elements were analyzed using the following standards:

GSP - granodiorite

BCR - basalt

AGV - andesite

G-2 - granite

PCC - peridotite

RGM - rhyolite

Trace elements were analyzed using the same standards but BBR (basalt) was substituted for BCR. All of the volcanic rocks analyzed are classified as tholeiitic or calc-alkaline using the method of Irvine and Baragar (1971; Fig. 11); and this diagram also illustrates the transitional nature of some samples. Furthermore, some rocks plot close to the F-M line, indicating a stronger depletion in Na_2O and K_2O (Jolly 1980). Conversely, a few samples

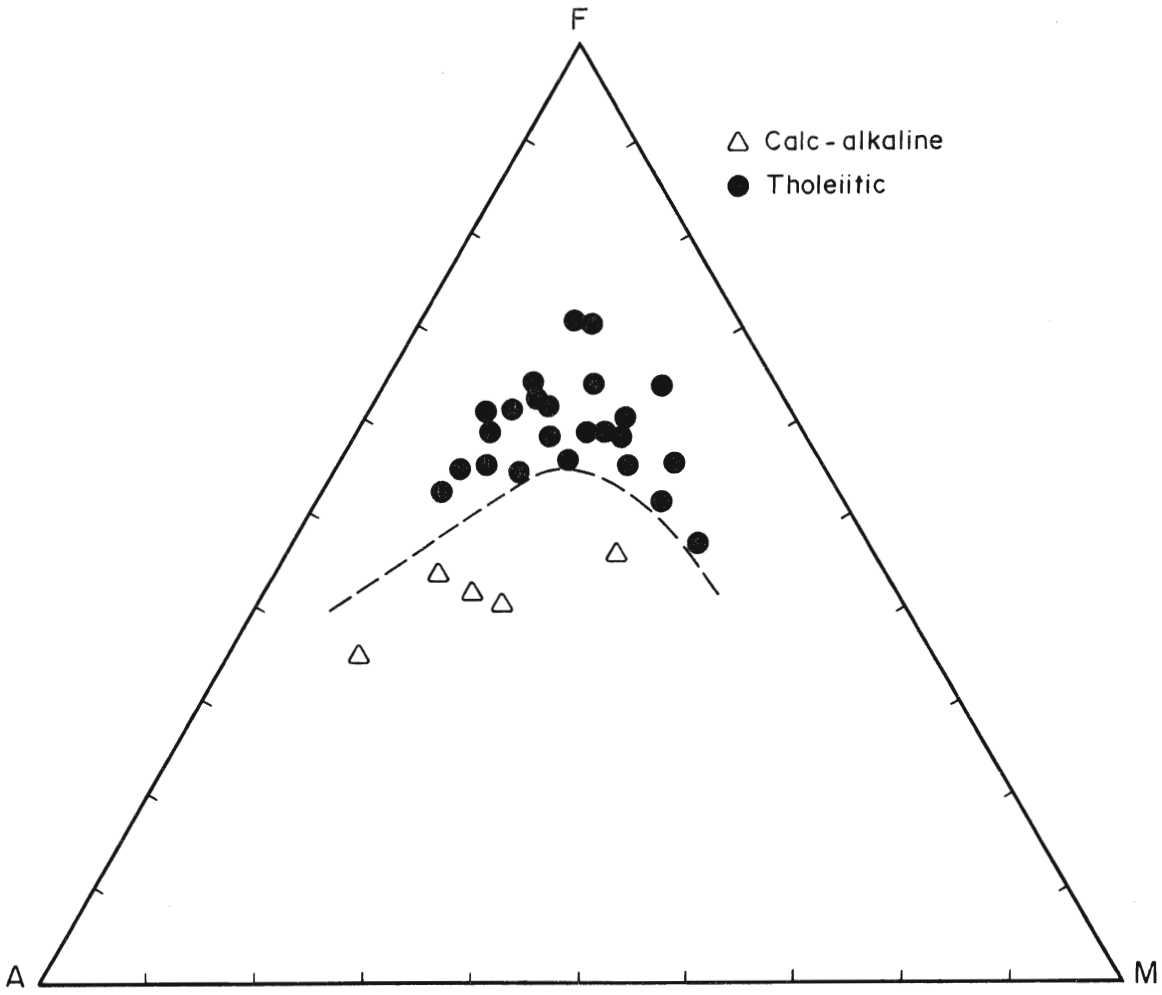

Figure 11. AFM diagram for Elliot Lake volcanic rocks (Irving and Baragar 1971).

plot close to the A-F line indicating an enrichment in alkali elements. These features are also evident in MgO-Na_2O and MgO-K_2O diagrams (Fig. 12). Tholeiitic and calc-alkaline rocks are sub-alkaline in nature, a relation illustrated by the total alkaline versus SiO_2 diagram. However, the highly mobile nature of Na_2O and K_2O during low grade metamorphism, suggest that alkali values of the analyzed specimens have been changed during alteration, thus this diagram has not been employed. Further distinctions between the suites can be gained from FeO/MgO-FeO, FeO/MgO-TiO_2 and FeO+Fe_2O_3-SiO_2 diagrams. These diagrams reflect the iron and titanium enrichment of tholeiites attributed to fractional crystallization (Osborn 1959, 1962; Jolly 1980), and the depletion of iron and titanium that occurs with calc-alkaline magmas. The FeO+Fe_2O_3-SiO_2 diagram also illustrates a strong silica gap suggesting that tholeiitic volcanism ceased and changing chemical conditions led to the formation of a new magma type. The distinction between the volcanic classes is facilitated by the weight percent of silica. For the purposes of this study the following subdivisions based on silica content are used:

Ultra basic - 45

Basalts - 46-50

Basaltic Andesite - 50-55

Andesite - 56-62

THOLEIITIC ASSOCIATION

Most of the volcanic rocks sampled from the Elliot Lake region are tholeiites. These rocks are characterized by a distinct iron enrichment trend as illustrated in Figures 13 and 14. Furthermore, similar iron enrichment trends are reported by Card *et al* (1977), Stout and Nicholls (1977), Jolly (1980), Flagler (1985) and McNaughton (1985).

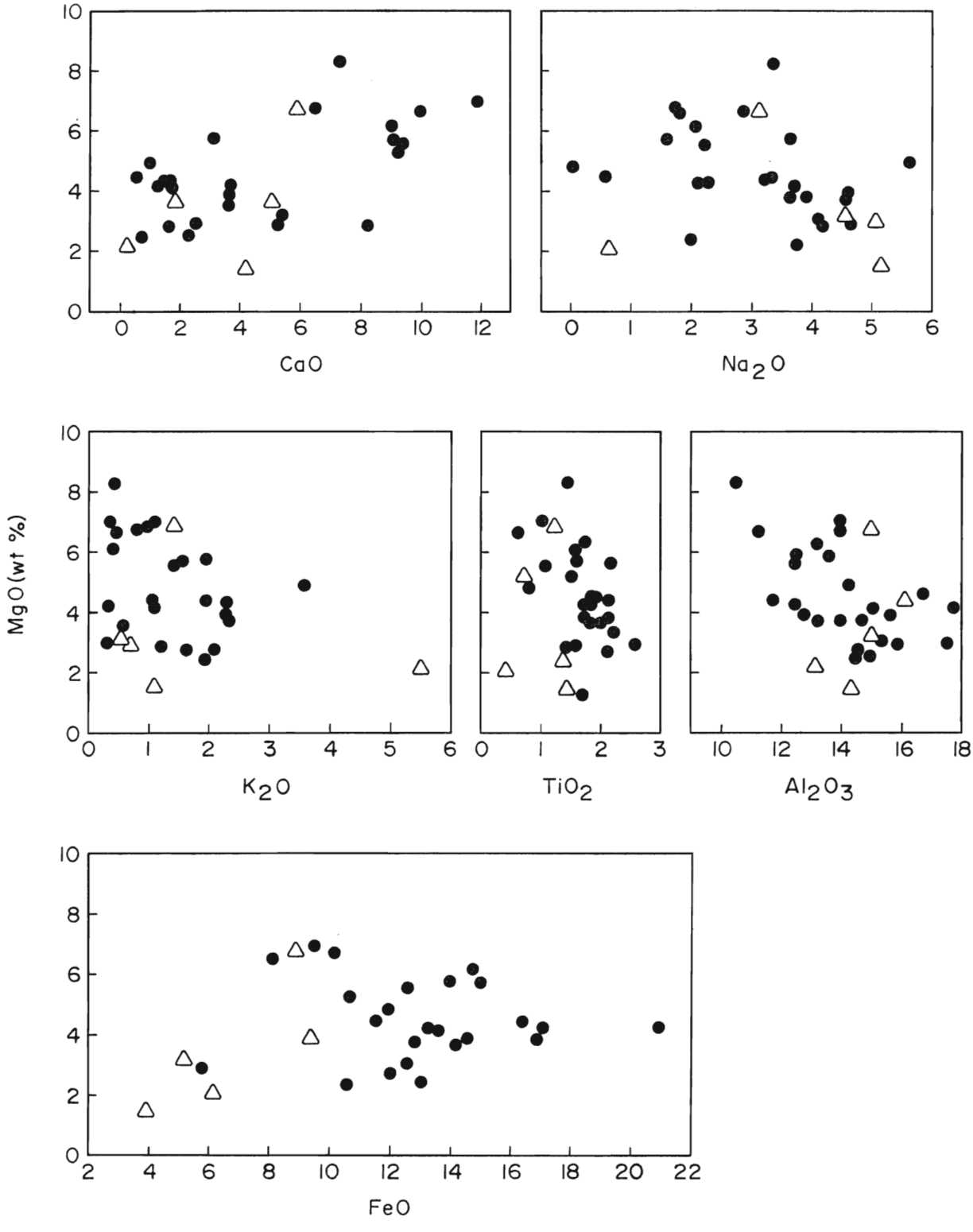

Figure 12. Variation diagrams of MgO versus major oxides in wt.%.

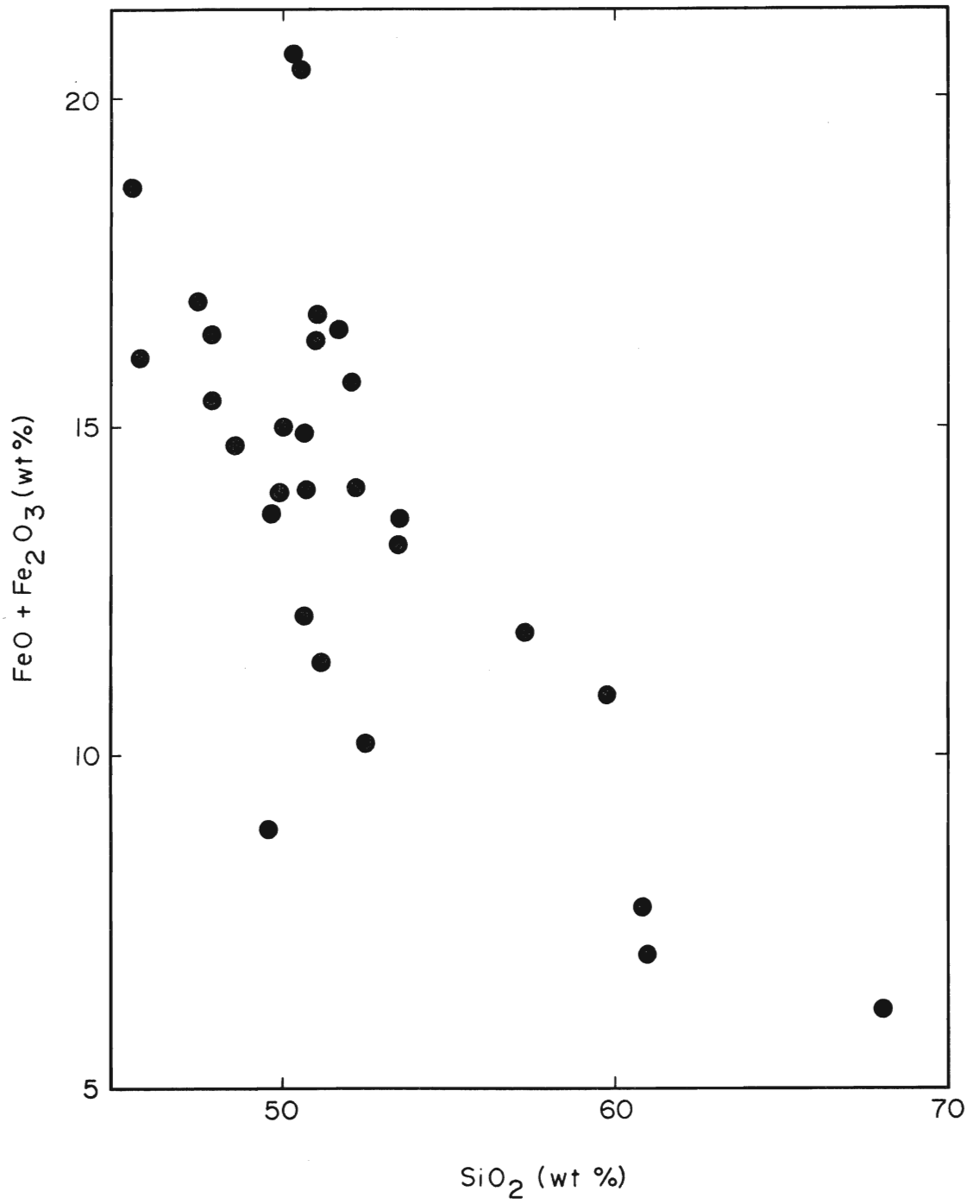

Figure 13. FeO + Fe_2O_3 - SiO_2 diagram for Elliot Lake volcanic rocks.

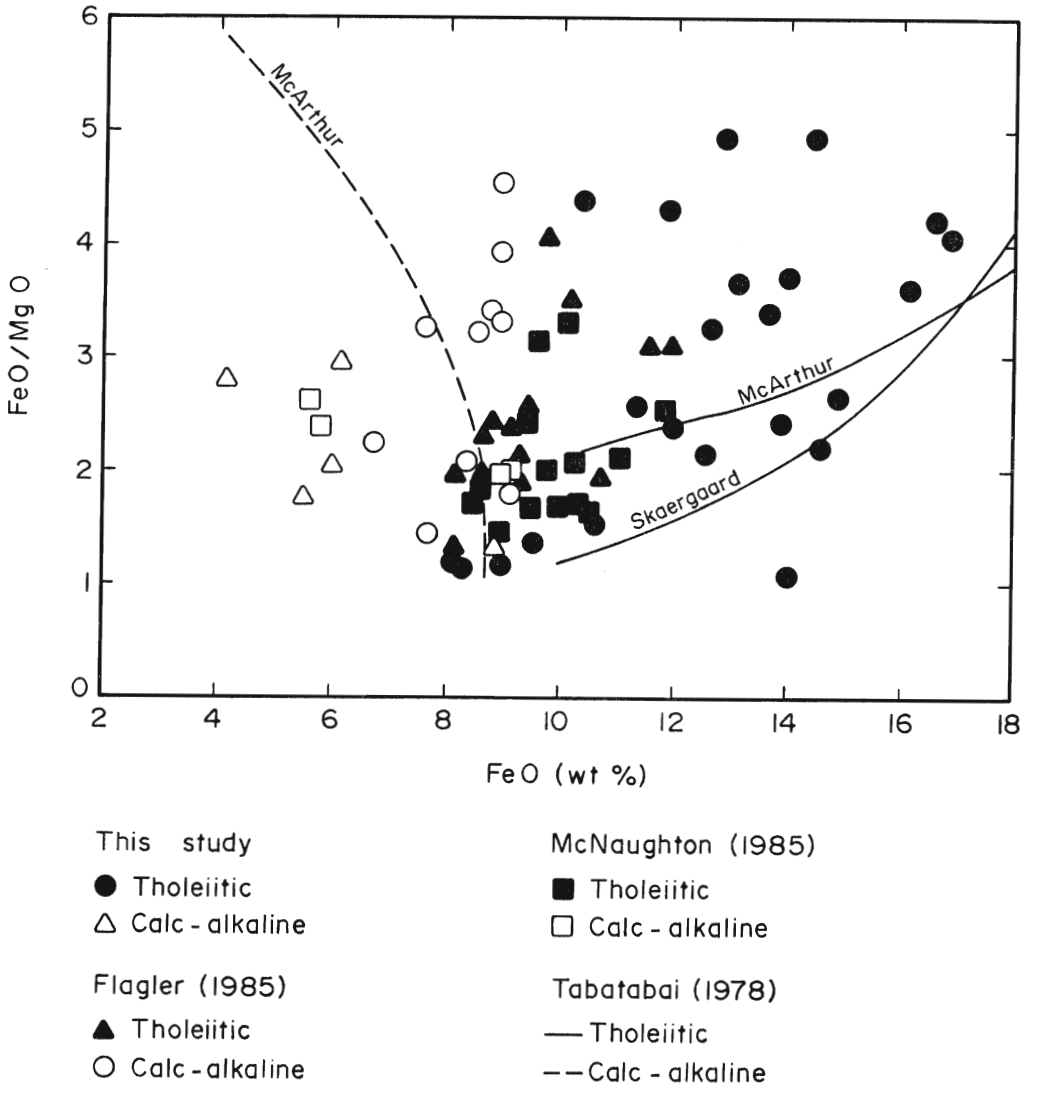

Figure 14. FeO/MgO-FeO wt.% diagram of tholeiitic and calc-alkaline rocks of the current and other studies.

Basalts and basaltic andesites are the dominant rocks present. These rocks show a definite TiO_2 enrichment trend with respect to silicia, due to an affinity of TiO_2 for the residual liquid during early fractional crystallization stages (Wood *et al* 1979). This is evident in MgO-TiO_2 and FeO-MgO-TiO_2 diagrams (Fig. 12 and 15). This trend parallels the TiO_2 enrichment trends of McArthur Township tholeiites and Skaergaard liquids (from Tabatabai 1978) (Fig. 15): and similar Thessalon volcanics described by Card *et al* (1977), Flagler (1985) and McNaughton (1985) and also concurs with the results of Clarke (1975). TiO_2 is an important parameter in distinguishing volcanic suites as outlined by Pearce and Cann (1973). TiO_2 values for Pecors Lake, Dollyberry Lake and Crazy Lake volcanics average 1.62 wt percent. However, basalts average 1.89 wt percent and basaltic andesites and andesites average 1.25 and 1.10 wt percent respectively. These values are higher than those reported by Tabatabai (1978), Card *et al* (1977), MacLean *et al* (1982), McNaughton (1985) and Flagler (1985), but are well below the results reported by Clarke (1975). However, similar TiO_2 values were cited by Jolly (1980) from Abitibi tholeiites which average 1.66 wt percent.

MgO variation diagrams express the relationship of other oxides to magnesium. MgO was used as a reference due to the fact that this oxide undergoes continuous depletion during fractional crystallization (Jolly 1980).

Aluminum oxide values for tholeiitic rocks average 14.11 wt percent which parallels the values reported by Clarke (1975) and Jolly (1980); but is less than the results of MacLean *et al* (1982). Similar values were established by McNaughton (1985) and Flagler (1985).

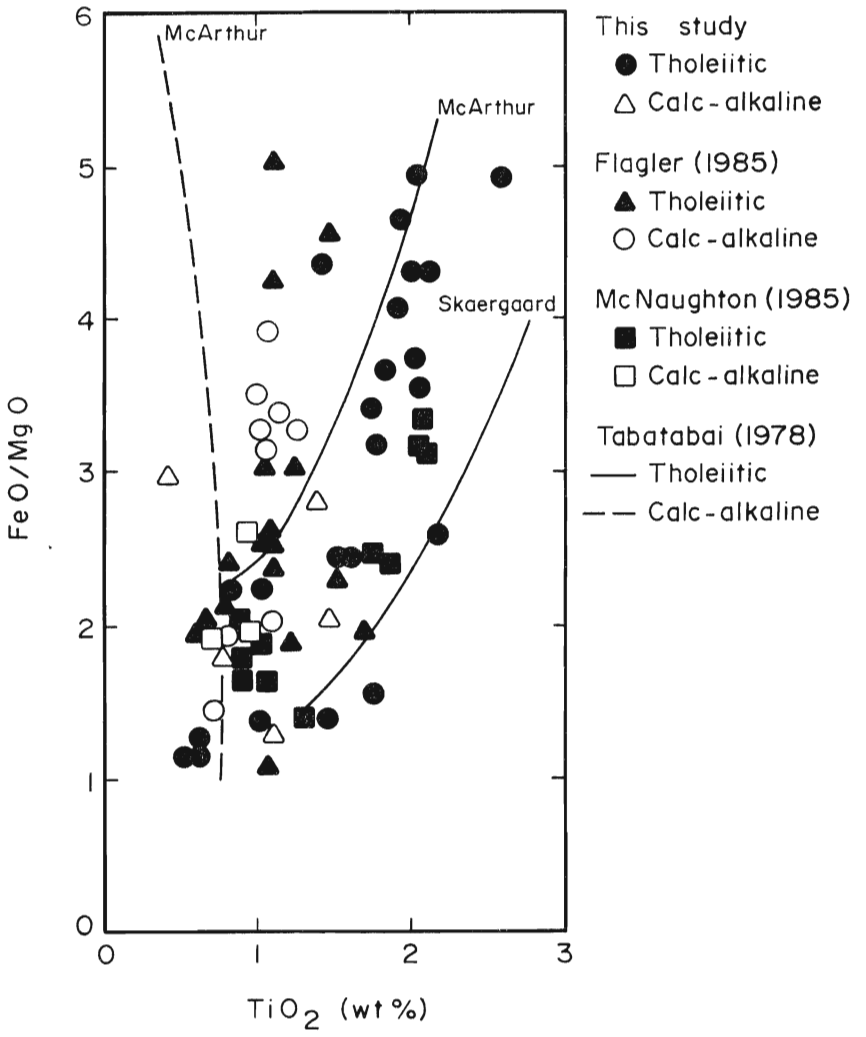

Figure 15. FeO/MgO-TiO_2 for tholeiitic and calc-alkaline rocks of the current and other studies.

An enrichment trend is observed for iron oxide. Iron is discussed in this section on the basis of total Fe as FeO, because as a result of metamorphic alterations oxidation states of the lavas do not reflect original compositions (Jolly 1980). A tholeiitic iron enrichment trend was also reported by Clarke (1975), Card *et al* (1977), Stout and Nicholls (1977), Jolly (1980), McNaughton (1985) and Flagler (1985). The rocks of the present study show a greater enrichment in iron than other Thessalon volcanics but are similar to the FeO enrichment of Archean tholeiites studies by Tabatabai (1978) and Jolly (1980).

MgO-CaO, MgO-Na_2O, MgO-K_2O variation diagrams (Fig. 12) reveal that no definite trends exist. Therefore there seems to be no relationship between CaO, Na_2O and K_2O values and fractional crystallization. These oxides are highly mobile during low grade metamorphism (Jolly 1980; Suk 1983) and hence the values do not reflect fractionation. The average CaO value for Elliot Lake tholeiites is 4.70 wt percent. This figure is low compared to other Huronian tholeiite values reported by Card *et al* (1977), McNaughton (1985), and Flagler (1985), but well below the CaO values established by Clarke (1975), Jolly (1980), and MacLean (1982). Elliot Lake Thessalon volcanics are low in K_2O, with tholeiites averaging 1.59 wt percent which is greater than the values reported by Card *et al* (1977); Jolly (1980), McNaughton (1985), Flagler (1985) and MacLean *et al* (1982).

Sodium oxide values for tholeiites average 2.83 wt percent which is the same value reported by Flagler (1985), lower than the values cited by Clarke (1975), Jolly (1980) and McNaughton (1985), but higher than the values reported by Card *et al* (1977) and MacLean *et al* (1982).

CALC-ALKALINE ASSOCIATION

The calc-alkaline suite is composed of basalts and andesites. These rocks are differentiated from tholeiites by a depletion of iron and titanium throughout fractional crystallization (Osborn 1959, 1962; Jolly 1980; Dostal *et al* 1982). This is compensated by the increase of alkaline elements and silicon.

Depletion in iron is examplified by low FeO values. FeO averages 6.09 wt percent which is well below 6.64, 8.34, 7.93 and 9.15 wt percent reported by Tabatabai (1978), Jolly (1980), McNaughton (1985), and Flagler (1985). A similar depletion is illustrated by TiO_2 indicated by the MgO-TiO_2 diagram. Calc-alkaline rocks average 1.06 wt percent and this value parallels the results reported by Jolly (1980) and Flagler (1985), but is higher than TiO_2 concentrations of Dostal *et al* (1982), Tarney *et al* (1982) and McNaughton (1985) which are 0.77, 0.92 and 0.87 wt percent respectively.

Aluminum oxide values are relatively constant compared to their tholeiitic equivalents. Elliot Lake calc-alkaline rocks average 14.60 wt percent which is greater than the 13.52 and 11.37 obtained by McNaughton (1985) and Flagler (1985) but lower than the 15.5, 15.62 and 17.6 wt percent reported by Jolly (1980), Tarney *et al* (1982) and Dostal *et al* (1982).

A similar trend exists for CaO as Elliot Lake rocks exceed the CaO concentration of other Thessalon volcanics but do not exceed the values reported by Jolly (1980), Tarney *et al* (1982) and Dostal *et al* (1982).

Depletion of MgO is great as MgO averages 3.25 wt percent which is greater than Thessalon volcanics studies by Flagler (1985), but well below the values reported by Tabatabai (1978), Dostal *et al* (1982) and Tarney *et*

al (1982). Na_2O averages 3.9 wt percent which exceeds the values reported by Tarney *et al* (1982), Dostal *et al* (1982) and Jolly (1980). K_2O values average 1.80 wt percent for Elliot Lake volcanics which is less than the values of Tarney *et al* (1982) but greater than the K_2O concentrations of other Thessalon Formation calc-alkaline rocks examined by Flagler (1985) and McNaughton (1985).

TRACE ELEMENT CHEMISTRY

Trace element data for Elliot Lake volcanic rocks are presented in Appendix 1. MgO variation diagrams (Fig. 16 and 17) reveal a significant increase in Ba and Rb contents with decreasing MgO, and this is reflected by the increase in Ba from 693 to 835 ppm in tholeiitic and calc-alkaline rocks respectively. This enrichment of Ba and Rb parallels the results reported by Tabatabai (1978) and those of Flagler and McNaughton (1985). Huronian volcanic rocks from the Sudbury-Espanola region reported by Card *et al* (1977) contain Ba concentrations from approximately 100 to 500 ppm. These values are much lower than the results of the present study, but are significantly greater than results reported by Clarke (1975) and MacLean *et al* (1982). Values cited by all of these authors are much lower than Elliot Lake rocks, but Ba and Rb concentrations do increase in the transition from tholeiitic to calc-alkaline rocks.

However, these elements are not good indicators of fractionation trends as they can be highly mobile during metamorphism. Zr and Y illustrate very little change in concentration in response to decreasing MgO. Tholeiites average 124 ppm Zr and 13 ppm Y and calc-alkaline rocks average 130 ppm Zr and 10 ppm Y. These values are slightly lower than the ones reported by Flagler (1985), McNaughton (1985) and Card *et al* (1977), but

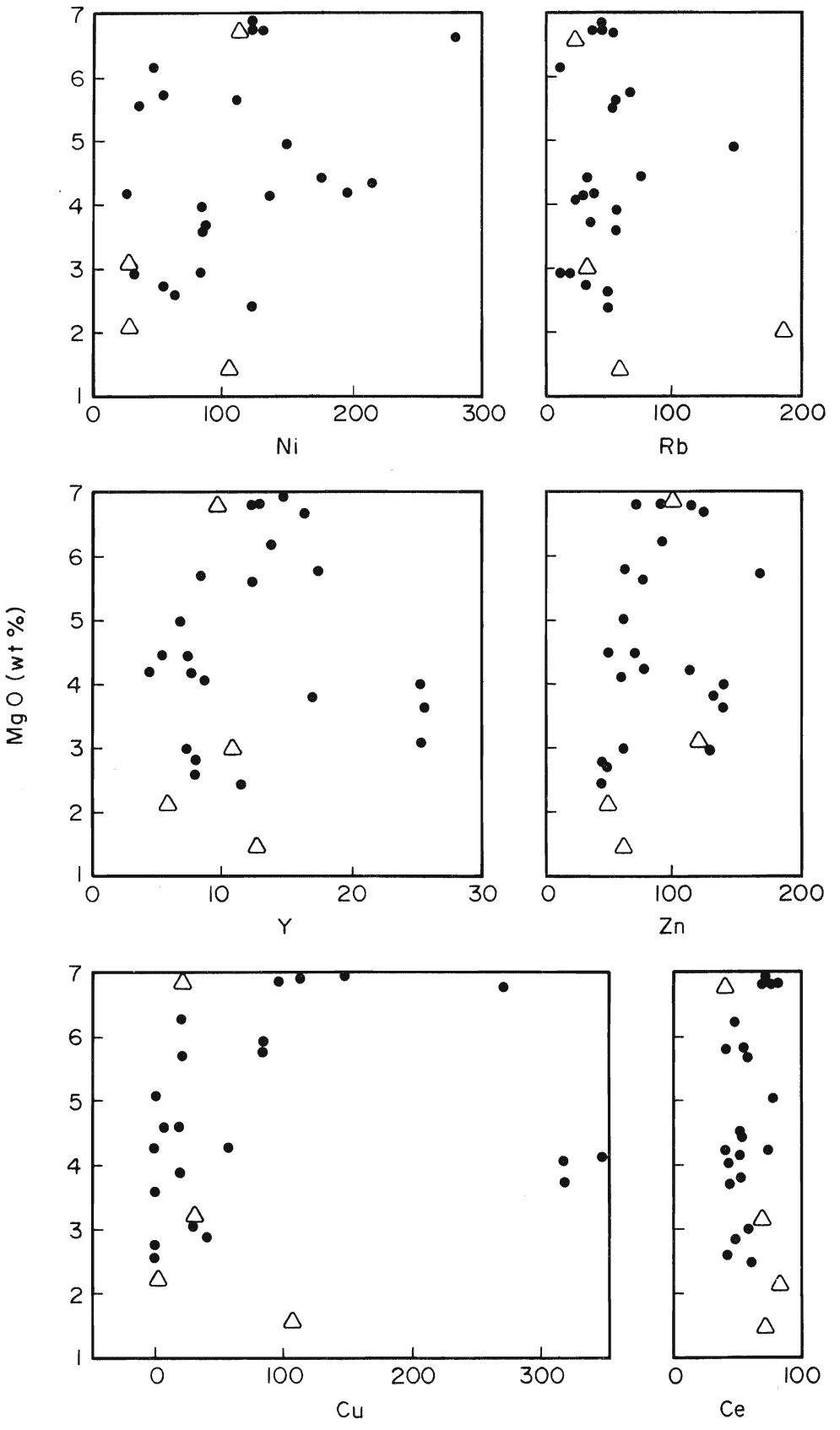

Figure 16. Variation diagrams of MgO versus trace element abundances in ppm.

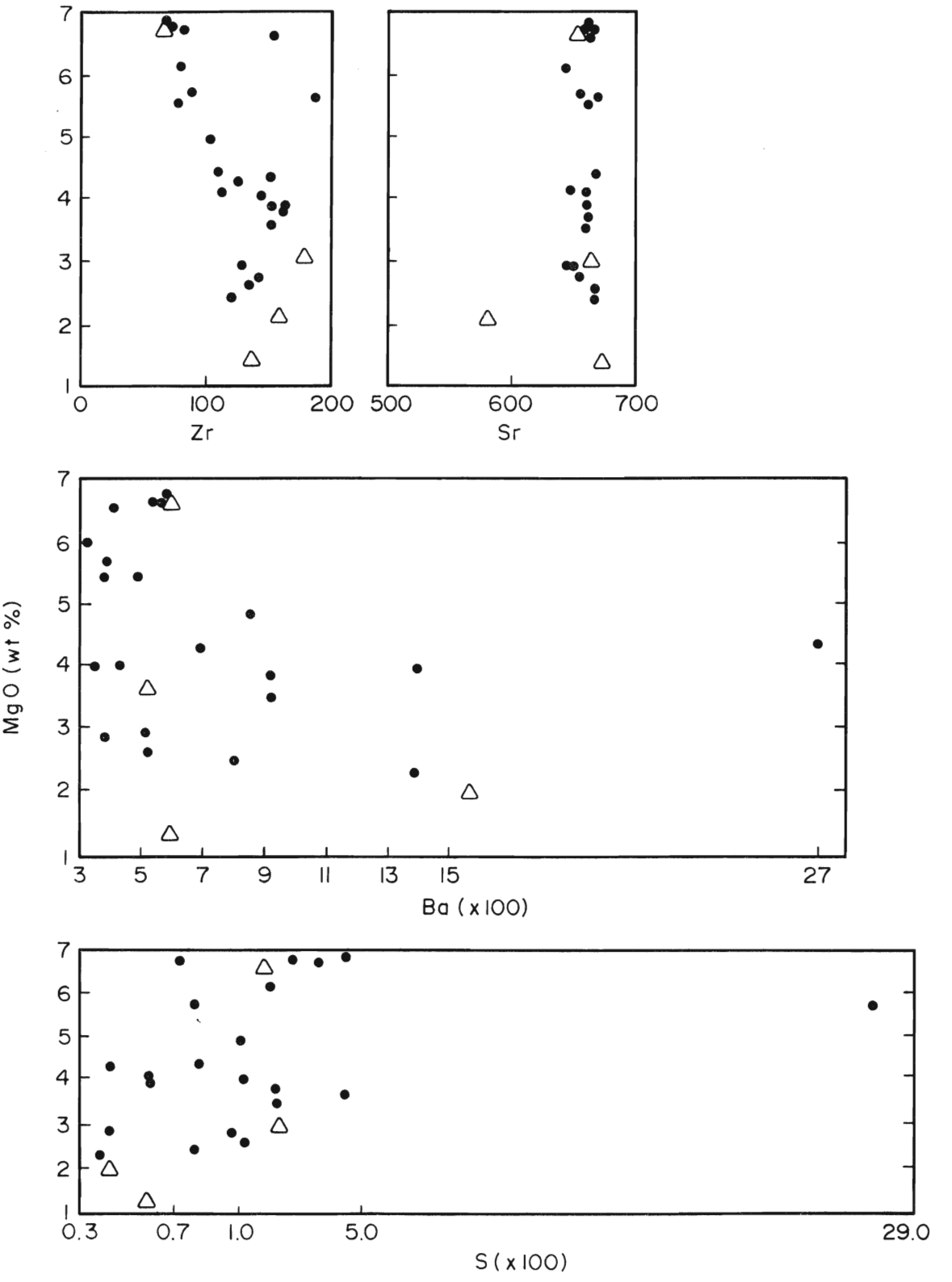

Figure 17. Variation diagrams of MgO versus trace element abundances in ppm.

are higher than values cited by Tabatabai (1978), Pearce and Norry (1979), MacLean *et al* (1982) and Davison (1984). Zr values are much lower than the results reported by Clarke (1975), for the metabasalts in his study average 182 ppm, which indicates a strong affinity of Zr for the liquid during fractionation processes (Wood *et al* 1979; Pearce and Norry 1979). However, all Thessalon Formation volcanics show very little change in Zr and Y concentrations between tholeiitic and calc-alkaline rocks. Y and Zr are concentrated in the residual liquid during fractional crystallization and fresh basalts show a range in concentration depending upon the amount of crystallization before eruption (Humphris and Thompson 1977). These elements show little concentration changes during alteration and can be used in addition to Ti for the identification of the original tectonic environments of volcanic rocks (Pearce and Cann 1973; Humphris and Thompson 1977; Pearce and Norry 1979). Sr values are much higher than the values reported by Flagler (1985), McNaughton (1985), Clarke (1975), Card *et al* (1977), MacLean *et al* (1982). The concentration of Sr decreases slightly with decreasing MgO (Fig. 17) but an average difference of 20 ppm between tholeiitic and calc-alkaline rocks is negligible. This observation is reported by Flagler (1985), Card *et al* (1977) and a difference of 2 ppm was obtained by McNaughton (1985). Sr is not a good indicator of fractionation trends as this element can be highly mobile during metamorphism (Pearce and Cann 1973; Smith and Smith 1976; MacLean *et al* 1982). Similar results were observed with Zn concentrations, for Zn exhibits very little change in response to a depleting MgO content.

S is abundant in Elliot Lake volcanic rocks but varies considerably from sample to sample. This is reflected by concentrations ranging from

37 to 2726 ppm. The average value of 283 ppm coincides with the 284 ppm reported by Tabatabai (1978).

The concentration of Ni ranges from 24.5 to 214.0 ppm in volcanic rocks of the present study. Tholeiitic rocks average 106.15 ppm Ni, and this value exceeds the Ni concentrations reported by Tabatabai (1978), McNaughton (1985), Clarke (1975), Card *et al* (1977), and MacLean *et al* (1982), but is lower than the value determined by Flagler (1985). The concentration of Ni decreases to an average of 88.2 ppm in calc-alkaline rocks from the Elliot Lake region, and a similar depletion occurs in other Thessalon Formation volcanics.

Ce generally shows very little change in concentration with changes in the MgO content as illustrated in Figure (16), but the concentration of this element is slightly greater in calc-alkaline rocks than tholeiites.

The concentration of Cu ranges from 0 to 343 ppm in Elliot Lake volcanic rocks. Many of the samples contain no copper as illustrated in Figure (16) and similar results were reported by McNaughton (1985) and Clarke (1975). The variation in the Cu concentration may be the result of alteration, as Cu is susceptible to alteration and is often leached into the circulating fluids (Humphris and Thompson 1977).

DISCRIMINATION DIAGRAMS

Many trace elements become mobile during metamorphism and do not give an accurate account of element enrichment or depletion during fractional crystallization (Pearce and Cann 1973; Smith and Smith 1976; MacLean *et al* 1982). However, elements such as Ni, Zr and Y are insensitive to metamorphic reactions and can be used with Ti to determine a tectonic environment (Pearce and Cann 1972; Humphris and Thompson 1977; Wood *et al* 1979). Y and Zr are concentrated in the residual liquid

during fractional crystallization (Humphris and Thompson 1977; Pearce and Norry 1979) while Ni is sensitive to olivine fractionation (Pearce and Cann 1973); Humphris and Thompson 1977). Ti is utilized in Ti-Zr-Y and Ti-Zr-Ni diagrams because it is unaffected by metamorphism. It illustrates enrichment in tholeiitic rocks and gradually diminishes in concentration during fractional crystallization. The ZTN Zr-Ti-Ni (Fig. 18) or the Zr-Y-N ternary diagram discriminates calc-alkaline rocks from tholeiites. In this manner, the ZTN diagram is similar to an AFM diagram. However, a comparison of the AFM and ZTN diagrams reveals a slight discrepancy, for the former produces fewer calc-alkaline types than the ZTN diagram. Volcanic rocks established as tholeiites on the AFM diagram, but determined as calc-alkaline rocks on the ZTN diagram occur close to the line separating calc-alkaline from tholeiitic rocks on the AFM diagram (Fig. 11), due to their generally alkali-poor and iron-rich nature and are considered as tholeiites in this study. Calc-alkaline rocks determined from the AFM diagram plot below the tholeiitic trend line in the ZTN diagram and fall within the calc-alkaline field.

Similar results were cited by Flagler (1985), who studied volcanic rocks that were classified as tholeiite and calc-alkaline types. An enrichment in Zr occurred in many rocks with a complementary depletion in Ni, and this is indicative of a higher degree of fractionation (Pearce and Cann 1973; Humphris and Thompson 1977). The ZTN diagram shows tholeiite rocks and many calc-alkaline types including the rhyolites reported by Weichelt (1984).

To determine the tectonic environment of the Elliot Lake volcanic rocks a discrimination diagram using elements of low mobility, Ti, Zr and Y are used (Fig. 19). This method developed by Pearce and Cann (1973)

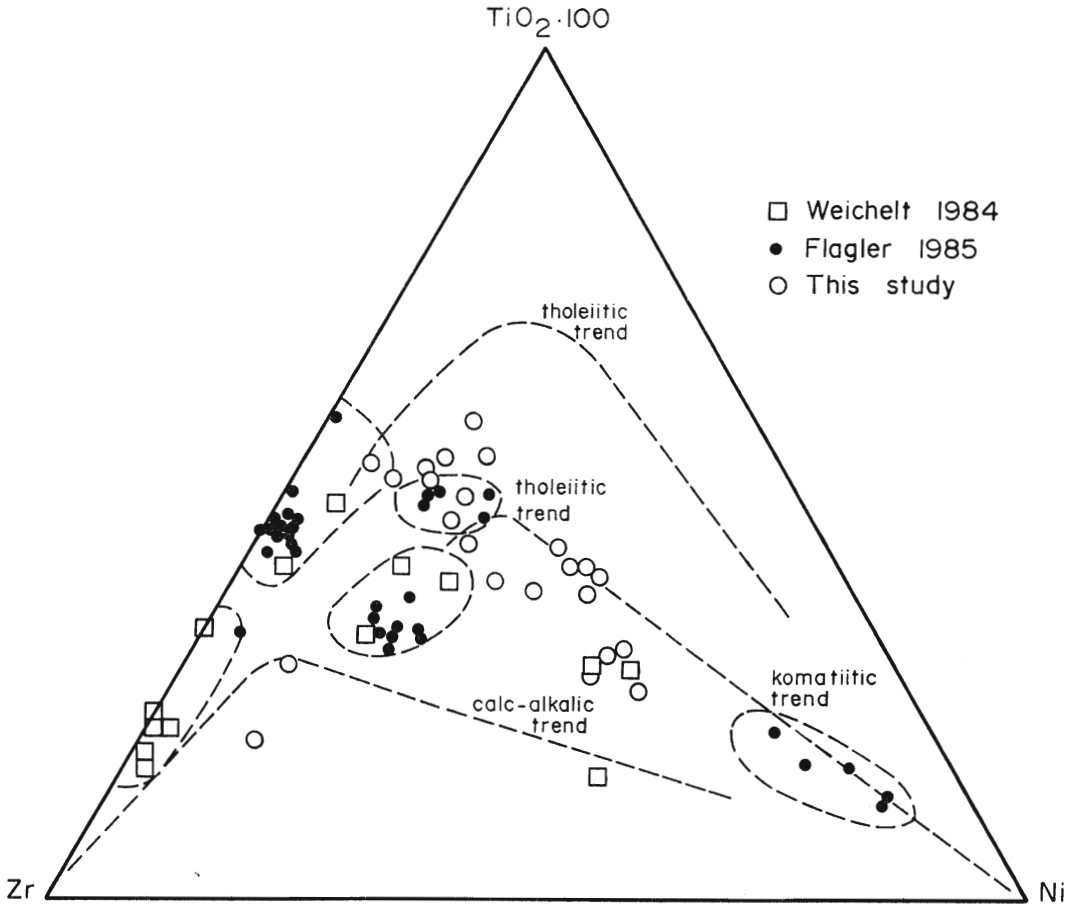

Figure 18. TiO_2 x 100-Zr-Ni diagram for the volcanic rocks from the Elliot Lake region of northern Ontario (Tabatabai 1978).

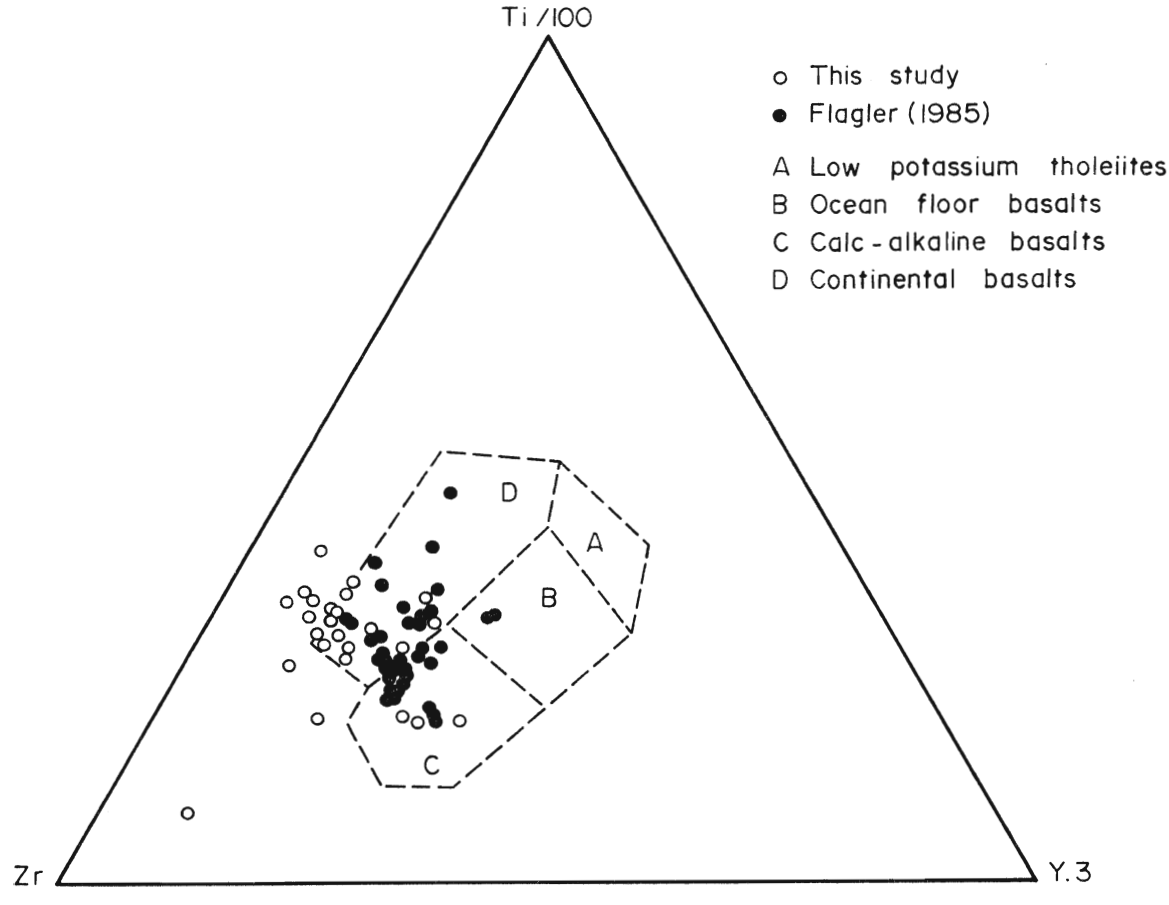

Figure 19. TiO_2/100-Zr-Yx3 ternary diagram for volcanic rocks of the Elliot Lake region of northern Ontario.

differentiates low-K tholeiites, calc-alkaline basalts, continental basalts and ocean-floor basalts. The data of Thessalon Formation volcanic rocks from the Elliot Lake region fall within the continental ocean floor basalt field (Fig. 19). Many samples plot outside all defined fields of the Ti-Zr-Y plot. The lines marking the different fields are quite arbitrary and since the unclassified samples plot close to field D, they will be considered continental rocks.

Most results obtained by Flagler (1985) for Thessalon volcanic rocks plot within the fields of continental basalts and calc-alkaline basalts (Fig. 19), and similar observations were reported by McNaughton (1985) and Card *et al* (1977).

DISCUSSION

Low grade metamorphism has destroyed primary igneous textures and degraded all primary mineralogy, including plagioclase, which has been extremely albitized. The presence of chlorite, albite, epidote and actinolite places these rocks within the greenschist facies, with a temperature range of 325°C (Schiffman and Liou 1980) and 450°C (Studemeister 1983). Metamorphism has led to the mobility of alkali elements. Immobile major and minor element oxides are used to determine the origin of the Elliot Lake tholeiitic and calc-alkaline lavas as these elements illustrate the effects of fractional crystallization.

Tholeiitic lavas normally arrive at the earth's surface having been produced from basic liquids with temperatures of 1250°C (Huppert *et al* 1984). This basic magma results from a low degree of partial melting (5-20%) of mantle material (Green 1975; MacLean *et al* 1982) due to local increases in temperature (Cox *et al* 1979).

The dominant tholeiitic rock in the Elliot Lake area is basalt; these are normally aphanitic - microporphyritic. Basalts reflect fractional crystallization at approximately 10 kb or 30 km (Cox and Horving 1966; Arth and Hanson 1975) while calc-alkaline suites may represent partial melting of a parental basaltic magma (Osborn 1959, 1962; Green and Ringwood 1968; Green 1975; Tarney *et al* 1982). Calc-alkaline rocks may also form by the replenishment of source rocks in the mantle from descending lithospheric slabs (Windley 1979; Young 1981; Condie 1982). Partial melting of mafic crust provides the necessary water concentrations needed to form calc-alkaline rocks (Green and Ringwood 1968; Condie 1982). Under high PH_2O water conditions, oxygen fugacity is raised (Osborn 1959, 1962; Moore 1979) and iron oxide minerals will form. Both of these mechanisms for the formation of calc-alkaline rocks result in a magma depleted in iron and titanium and enriched in silica. This is evident in the $FeO+Fe_2O_3$-Si_2 and FeO/MgO-TiO_2 diagrams (Figs. 13 and 15).

A decrease in iron is consistent with the appearance of magnetite as suggested by Osborn (1959, 1962). Under high oxygen fugacity, magnetite crystallizes leading to a depletion of Fe and Ti minerals in the magma. Thus, fractional crystallization can produce two trends resulting from changes in liquid composition (Osborn 1959, 1962). The composition of the mixture can remain constant and pO_2 can stay constant or increase, and the resulting liquid composition changes with the increase in oxygen. Therefore, the pO_2 determines whether liquids move toward higher silica and lower iron oxide content, or toward a higher iron oxide content with little change in silica. Calc-alkaline rocks are representative of the second trend maintaining a constant (or increasing) pO_2. This characteristic of oxygen fugacity is due mainly to higher water content in magmas during

differentiation (Osborn 1959, 1962; Green 1975). Water is present in sufficient quantities in a magma reservoir, and in addition to slow cooling and crystallization pO_2 was maintained supplying 0.2% O_2 (Osborn 1962) during the change from basalts to more silica rich suites. The concomitant depletion of iron results from iron in the liquid being oxidized from Fe^{2+} to Fe^{3+}. Conversely, tholeiites occur with a decrease in pO_2 during fractional crystallization and the resulting liquid becomes rich in FeO and contains little H_2O (Osborn 1959, 1962; MacPherson, 1984).

Muenow *et al.* (1979) reported that tholeiitic basalts contain approximately 0.002 wt. % H_2O at the time of partial melting. Fractional crystallization causes H_2O to be concentrated in residual liquid and gas, and this is aided in part by the capacity of silicate melts to be water collectors (Harris 1981).

The presence of vesicles sheds some light on the amount of H_2O and CO_2 trapped in magmas. As already stated, tholeiitic magmas contain little or no H_2O. MacPherson (1984) reported that tholeiitic magmas contain initially H_2O in the range of 0.2-0.5 wt. % and 0.2 wt. % CO_2 and fewer vesicles than more alkaline rocks. Shallower eruptions produce more vesicular rocks and contain about 0.5 wt. % H_2O and low initial CO_2 contents of approximately 0.05 wt. % (MacPherson 1984). The exsolution of volatiles from magma during its ascent and eruption allows for the escape of gas, due to the decrease in confining pressure. Shallow eruptions produce more and larger vesicles because exsolved volatiles expand with lower pressure and exsolution of volatiles is more complete (Moore 1979).

The increase in H_2O content of the more alkaline rich magma at shallower depths can be related to calc-alkaline rocks fractionating at shallower depths than their tholeiitic parents (Osborn 1959,1962). These

calc-alkaline lavas contain more H_2O due to the increase in oxygen fugacity and silica content of the magma. Increased silica in more alkaline magmas is evident from the FeO + Fe_2O_3 - SiO_2 diagram (Fig. 13) and calc-alkaline rocks illustrate the greatest silica enrichment.

Thessalon Formation tholeiites contain fewer vesicles than the calc-alkaline rocks. This indicates that the tholeiitic magmas contain less H_2O than calc-alkaline magmas (Muenow *et al.* 1979; MacPherson 1984) and that the presence of vesicles is indicative of gas exsolution (Osborn 1959, 1962; Moore 1979). Vesicle formation may be attributed to other factors such as melt viscosity. Increasing viscosity of a melt inhibits degassing (Anderson *et al.* 1983) while less viscous melts allow magmas to degas more quickly . The presence of amygdules in close association with vesicles may indicate a genetic relationship. Some authors report that deformed amygdules represent flow structures (Bennett 1976), but no such features were observed during this study. Therefore it may be reasonable to propose that amygdules also represent degassing structures.

The metamorphic assemblages for Elliot Lake Thessalon Formation volcanics are exhibited in the ACF diagram (Fig. 20). During greenschist metamorphism the alkali elements are highly mobile. Albitization of pre-existing plagioclase results in the liberation of Ca^{2+} ions, which then can be consumed to form calcium bearing minerals such as actinolite, sphene, epidote and calcite. However, the strong depletion of calcium in some samples as shown by the ACF diagram may reflect the lack of actinolite in these samples.

Epidote occurs as a groundmass constituent, a replacement feature in plagioclase or as a void filling and therefore is late entering the system. The presence of sphene represents a mechanism for the uptake of Ca^{2+} ions,

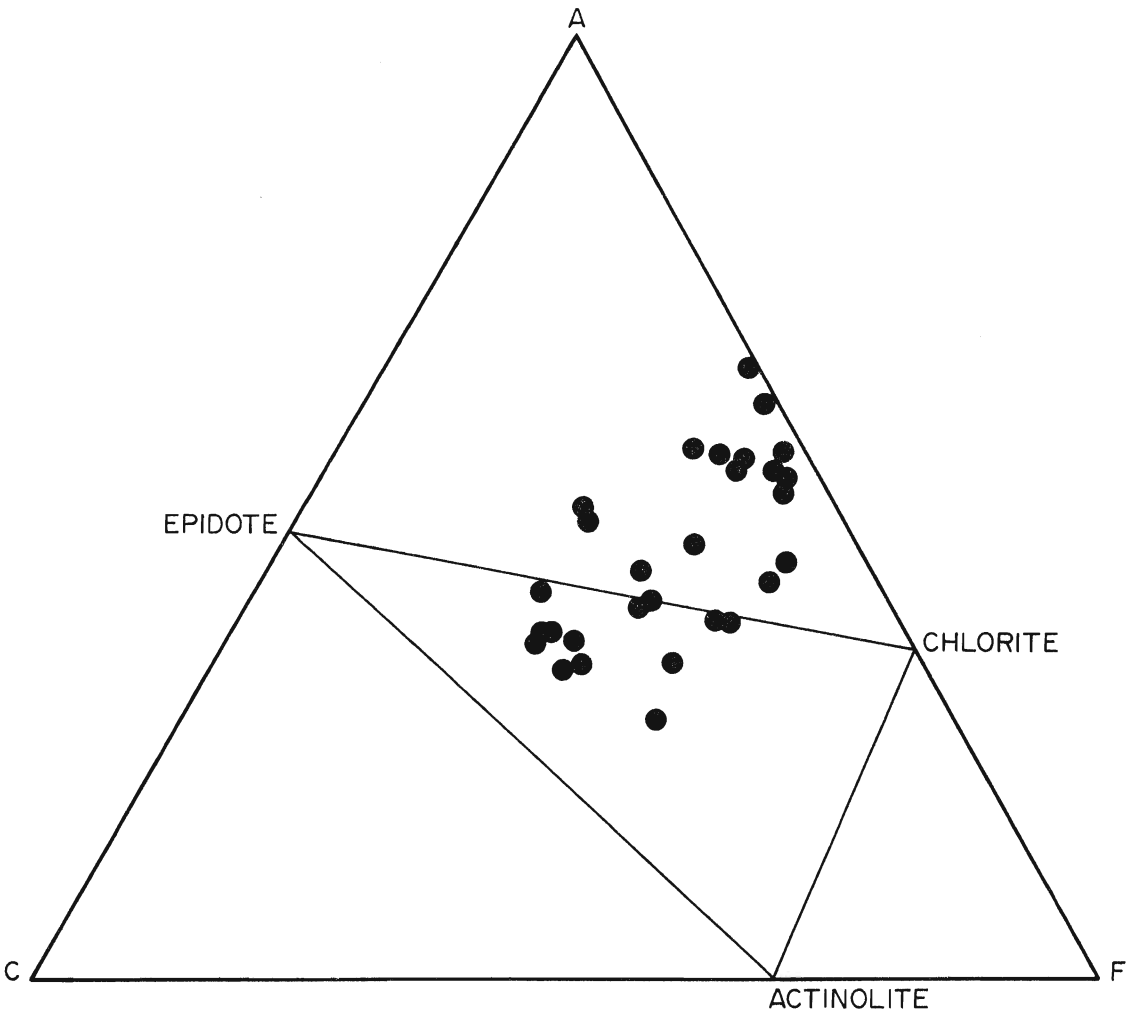

Figure 20. ACF diagram of the Elliot Lake region volcanic rocks.

and also indicates that CO_2 pressures were very low. The maximum CO_2 pressure falls in the range of 0.1 to 1.0 kPa (Shuiling and Vink 1967). Should the CO_2 pressure increase sphene would breakdown and rutile would appear.

All of the Elliot Lake volcanic rocks contain biotite, thus these samples represent the biotite zone of the greenschist facies (Winkler 1979). Ubiquitous biotite suggests considerable K_2O introduction by more alkaline rich intrusions (Jolly 1978).

Penetrative deformation occurs on a local scale, particularly near faults and other tectonic features. Petrographic observations reveal an east-west trending schistosity, which represents a penetrative deformation structure characteristic of regional metamorphic activity.

The lack of glass, pillow and quench textures and brecciated rocks suggests that these volcanic rocks do not represent submarine eruptions, but indicate that the bulk of the eruptions are subaerial in character (Clarke 1975). These igneous rocks are due to large amounts of lava flowing through fissures in the earth's crust which may be linked with rifting (Osborn 1962; Bennett 1976; Nisbet and Walker 1982). Pecors Lake, Dollyberry Lake and Crazy Lake volcanic belts lie along strike from one another, and these formations are situated parallel with the major faults in the Elliot Lake region, suggesting that the distribution of these volcanic rocks may be fault controlled.

Immobile element diagrams, particularly the Ti-Zr-Y diagram (Fig. 19) indicate that the Elliot Lake volcanics represent rocks from a continental environment (Pearce and Cann 1973). The moderately high Zr and Y concentrations parallel the Zr and Y contents of the within-plate basalts reported by Morrison (1978) and are higher than the Zr and Y

concentrations of island arc rocks established by MacLean *et al.* (1982) and Tarney *et al.* (1982). These elements show very little change in concentration during the transition from tholeiites to calc-alkaline rocks for Elliot Lake samples, as tholeiites average an increase of 6ppm Zr and Y decreases 3ppm, which are negligible changes in concentration. Zr does increase slightly during fractionation as evident from Figure 17, but begins to decrease at intermediate silica levels, and does not increase progressively with fractionation (Tarney *et al.* 1982).

The moderately high weight percent of TiO_2 in calc-alkaline rocks concurs with the values revealed by Morrison (1978), and are much greater than results obtained by Dostal *et al.* (1982) and Tarney *et al.* (1982). TiO_2 concentrations in tholeiites exceed the values cited by MacLean *et al.* (1982), whose rocks were established to be island arc or back arc basalts.

Ni shows similar results. Elliot Lake tholeiites contain much more Ni than the values reported by Card *et al.* (1977), MacLean *et al.* (1982) and Flagler (1985). The concentration of Ni in calc-alkaline rocks exceed the Ni content of the calc-alkaline rocks studied by Tabatabai (1978), Tarney *et al.* (1982) and Dostal *et al.* (1982).

The Ti-Zr-Ni diagram differentiates calc-alkaline rocks from tholeiitic samples, and reveals the general enrichment of Zr and TiO_2 in Elliot Lake volcanics and also illustrates the greater enrichment of Ni in Elliot Lake rocks than in the Thessalon Formation volcanics studied by Flagler (1985).

Other trace elements, such as Ce, Ba and Rb are not employed in discrimination diagrams but do reveal a relationship with silica as a result of fractionation. These elements increase in concentration with an increase in silica content and this trend is reported by Tarney *et al.* (1982),

Dostal *et al.* (1982) and MacLean *et al.* (1982) and previous workers studying other Thessalon Formation volcanics. However, Ba and Rb are mobile during metamorphism (Pearce and Cann 1973; Humphris and Thompson 1977; MacLean *et al.* 1982) and concentrations of these elements should be utilized with caution.

Elliot Lake volcanic rocks represent lavas produced in a continental environment. They lack pillow structures and brecciation, indicating that they were extruded subaerially (Clarke 1975). Volcanic lavas produced in a continental environment represent magmas erupted from fissures (Roscoe 1973; Turner *et al.* 1974) that are part of a continental rift system that is approximately 80 to 90 km wide and 1600 km long which represents a tectonic surface that produces high angle faults or a fault zone (Sims 1976; Chase and Gilmer 1973).

Tholeiitic rocks normally represent fissure flows that are the products of a basaltic liquid at low degrees of partial melting of a peridotite source (Green 1975 and MacLean *et al.* 1982). The progressive iron and titanium enrichment of this liquid is due to fractional crystallization (Osborn 1959, 1962; Jolly 1978).

Tholeiitic suites are succeeded by calc-alkaline suites which illustrate depleted iron and titanium contents, but indicate an increase in Si and Zr concentrations compared with tholeiitic rocks. Calc-alkaline volcanism may be the result of fractional crystallization of a basaltic magma at shallow depths in response to an increase in oxygen pressure (Osborn 1959, 1962; Clarke 1975; Jolly 1978), or these rocks may result from the partial melting of descending continental crust material (Tuttle and Bowen 19858; Windley 1979; Condie 1982). Partial melting of mafic crustal rocks provide the necessary water concentrations which are needed

to form calc-alkaline rocks (Green and Ringwood 1968; Condie 1982). The increased Zr concentrations of Elliot Lake calc-alkaline rocks seens to be the result of partial melting of crustal material, as descending lithospheric slabs lead to replenishment of silica and water (Condie 1982) and accounted for increased Zr concentrations (Tarney *et al.* 1982).

Other models have been proposed to account for the derivation of calc-alkaline rocks, and these include the partial melting of amphibolite or eclogite (Green and Ringwood 1968). Partial melting of eclogite under dry conditions leads to the formation of basaltic andesite and andesite. Conversely, if melting occurs under wet conditions, dacite or rhyodacite will form (Green and Ringwood 1968; Condie 1982).

Basaltic lavas with large water concentrations can recrystallize to amphibolite in lower parts of a basaltic successsion. Subsequent heating of this successsion may lead to partial melting of previously formed basalts. Magmas produced by this mechanism will not be water saturated, because very little water was present in the parent amphibolite.

These two processes account for an increase in silica content of the liquid by extracting aluminous amphibolite but do not account for the presence of iron oxides which are evident in Elliot Lake calc-alkaline rocks. Partial melting of eclogite produces calc-alkaline rocks chemically similar to those of the Elliot Lake region, but does not provide a mechanism to supply water essential for the formation of iron oxides and vesicles. Similarly, fractional crystallization of amphibolite will produce calc-alkaline rocks, but magmas producing calc-alkaline rocks contain little water (Green and Ringwood 1968).

Other mechanisms including direct partial melting of peridotite, assimilation of sialic crust and anatexis (Arth and Hanson 1975) can

produce intermediate to felsic rocks. However, these mechanisms fail to reveal a method to account for the formation of iron oxides, which indicate high oxygen and water concentrations characteristic of calc-alkaline magmas.

STATISTICS

The distinction between tholeiitic and calc-alkaline rocks is made on the basis of the concentrations of FeO and TiO_2. Tholeiites reveal a FeO and TiO_2 enrichment trend while calc-alkaline rocks are depleted in these elements. The enrichment in FeO and TiO_2 of tholeiites is illustrated in Figures 14 and 15. To support this observation, the weight percentage of FeO/MgO is plotted against FeO and TiO_2 in the scatter diagrams of Figures 21 and 22. Figure 21 represents the data for FeO/MgO-FeO, showing that weight percent FeO illustrates an enrichment trend. The equation of the line that best fits the data is $FeO/MgO=0.70\ FeO^{1.44}$ (Fig. 23, Table 4) which indicates a positive slope, and further supports an iron enrichment trend for tholeiitic rocks. This line, is the line of regression representing a power function. The power function line shows an enrichment in FeO weight percent in Figure 21 which parallels the trend illustrated in Figure 14. These two lines have the same concave upwards slope which is to be expected because both lines are plotted on arithmetic axes.

The correlation coefficient established for the power function FeO/MgO-FeO data is 0.74, a value indicating that a fairly strong relationship exists (Alder and Roessler 1968; Remington and Shork 1970). This correlation coefficient is the best value obtained as values of 0.65 and 0.70 for linear and exponential relationships do not indicate as strong an association between X and Y. The standard error of estimate can also be

Table 4. Statistical Data for Tholeiitic Rocks

DATA	REGRESSION LINE	CORRELATION COEFFICIENT		
		Power	Exponential	Arithmetic
FeO/MgO-FeO	$FeO/MgO=0.070FeO^{1.44}$	0.740	0.700	0.650
FeO/MgO-TiO_2	$FeO/Mg=1.875TiO_2^{0.898}$	0.795	0.788	0.749

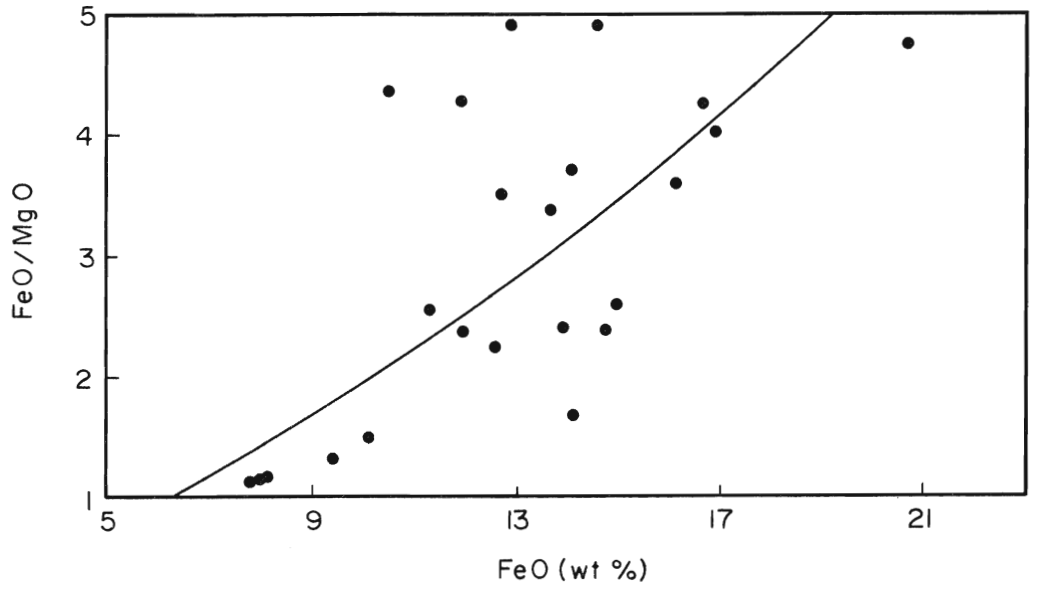

Figure 21. Scatter diagram and corresponding power curve for FeO/MgO-FeO data of tholeiitic rocks plotted on arithmetic axes(for statistics see Table 4).

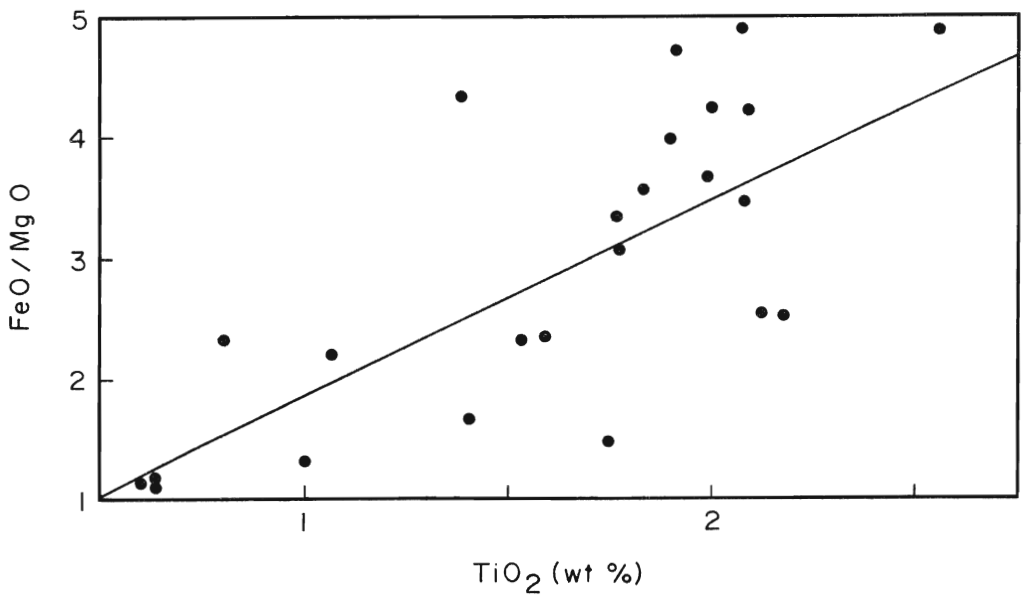

Figure 22. Scatter diagram and corresponding power curve for FeO/MgO-TiO_2 data of tholeiitic rocks plotted on arithmetic axes(for statistics see Table 4).

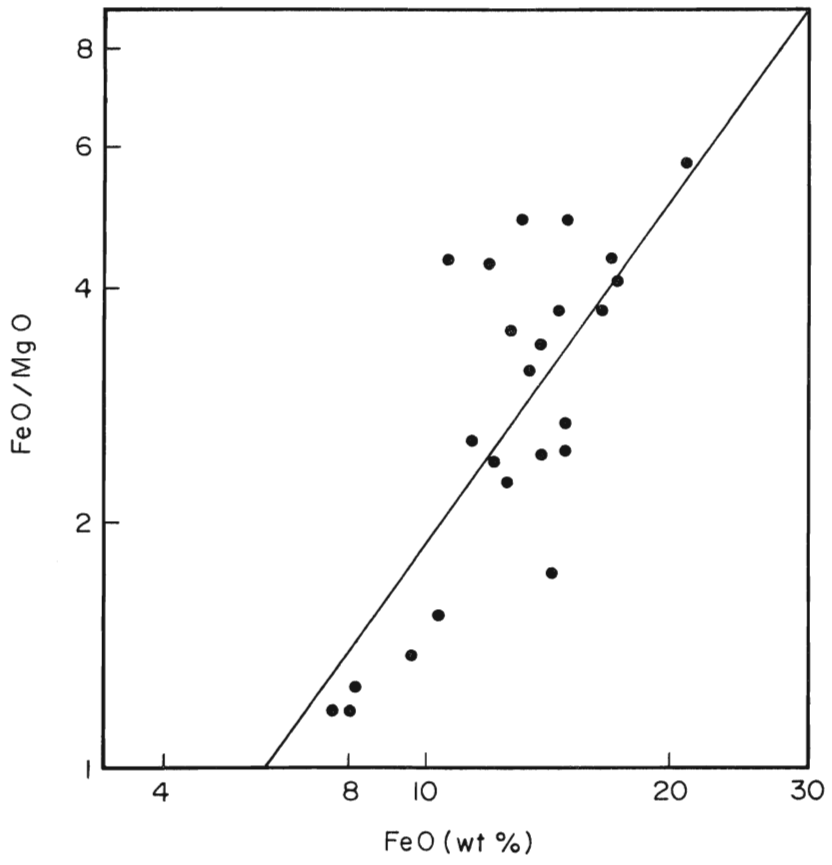

Figure 23. Regression line for FeO/MgO-FeO scatter diagram for tholeiitic rocks(for statistics see Table 4).

employed to determine the linear relationship of the data, as this value measures the spread of a set of points about a regression line (Remington and Shork 1970). The nearer the points to a straight line the smaller the standard error of estimate . The value of 0.143 indicates a small standard error of estimate for the data of the FeO enrichment trend, and thus supports the strong relationship established by the correlation coefficient. Figure 22 illustrates the data for FeO/MgO versus TiO_2. This scatter diagram reveals an enrichment trend of TiO_2 for tholeiitic rocks. The equation of the line that best fits the data is $FeO/MgO=1.875\ TiO_2^{0.898}$ (Fig. 24) and this line has a positive slope and this supports the observation that TiO_2 reveals an enrichment trend in tholeiitic rocks. As in the preceeding example, this line is the line of regression for the corresponding scatter points and represents a power function that also plots as a curved line on arithmetic axes (Fig. 22).

The correlation coefficient established from the FeO/MgO-TiO_2 data is 0.795. This correlation coefficient of FeO/MgO-TiO_2 data reveals the strongest relationship possible for the data because correlation coefficients established for arithmetic and exponential functions are 0.749 and 0.788 respectively.

The established regression lines for the FeO/MgO-FeO and TiO_2 data from Elliot Lake tholeiitic rocks illustrate an increase in concentration of FeO and TiO_2 weight percent and this parallels the results illustrated in Figures 14 and 15. The use of regression lines only supports the enrichment trends of FeO and TiO_2 in tholeiitic rocks as revealed in Figures 14 and 15. Tholeiitic rocks are discriminated from calc-alkaline rocks on the basis of FeO and TiO_2 concentrations and the use of statistical methods provides

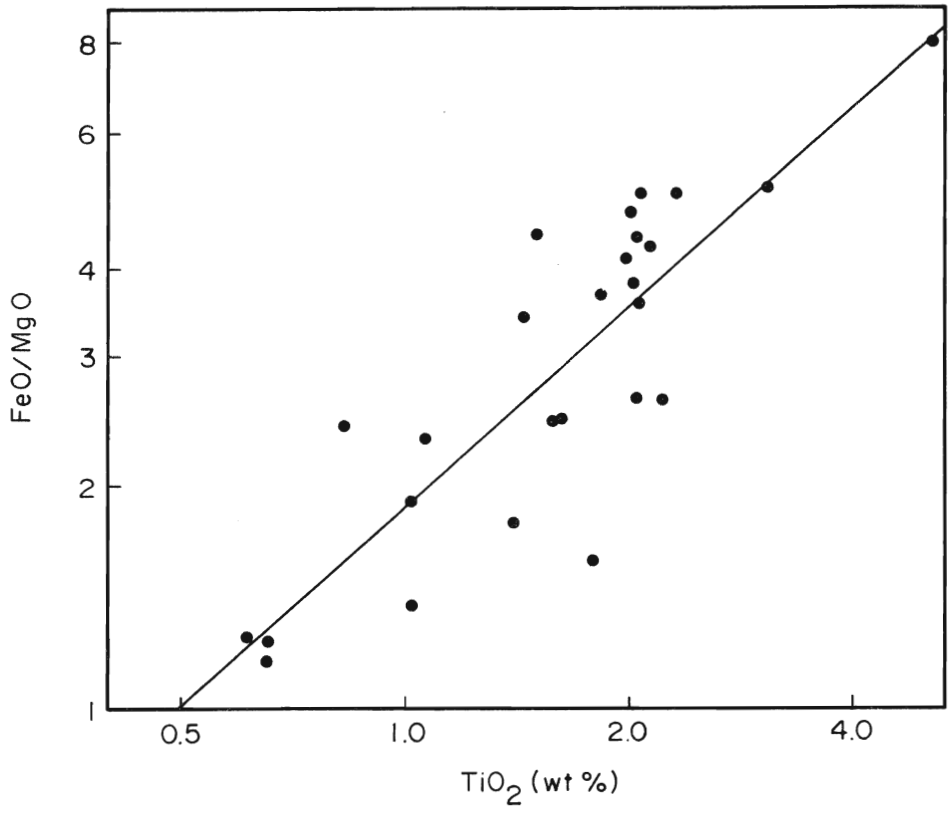

Figure 24. Regression line for FeO/MgO-TiO_2 scatter diagram for tholeiitic rocks(for statistics see Table 4).

support to the claim that tholeiitic rocks reveal a FeO and TiO_2 enrichment trend.

Similar scatter diagrams were attempted for calc-alkaline rocks, but only five calc-alkaline samples were available, hence the results obtained were not conclusive, and cannot be used to substantiate the FeO and TiO_2 depletion trends illustrated by Elliot Lake calc-alkaline volcanics.

TECTONIC MODEL

Thessalon Formation volcanics encountered in the Elliot Lake region were fissure fed to the earth's surface due to a continental rift system that is approximately 90 km wide and 1600 km long (Sims 1976; Chase and Gilmer 1973). The tectonics of the Huronian basin are not well understood (Roscoe 1970; Young 1981), but fissures appear to have resulted from large tectonic movements along the border of lower Precambrian granite and greenstone terrains which produce high angle faults or fault system (Fig. 25; Sims 1976). This border represents a zone of crustal weakness and provides a depression to trap accumulating volcanics and sediments.

Some authors, (Frarey and Roscoe 1970; Robertson 1972) reported that arkosic conglomerates are intercalated with volcanics in the lowermost layers of the Huronian Supergroup and this suggests that the site of volcanism may have been on the margin of a marine environment (Frarey and Roscoe 1970). Tholeiitic magmas result from the partial melting of a mantle peridotite (Cox and Hornung 1966; Turner *et al.* 1974; Green 1975) and fractional crystallization results in the familiar iron and titanium enrichment trends of tholeiitic magmas. This liquid has a density less than surrounding mantle material and this results in the diapiric rise

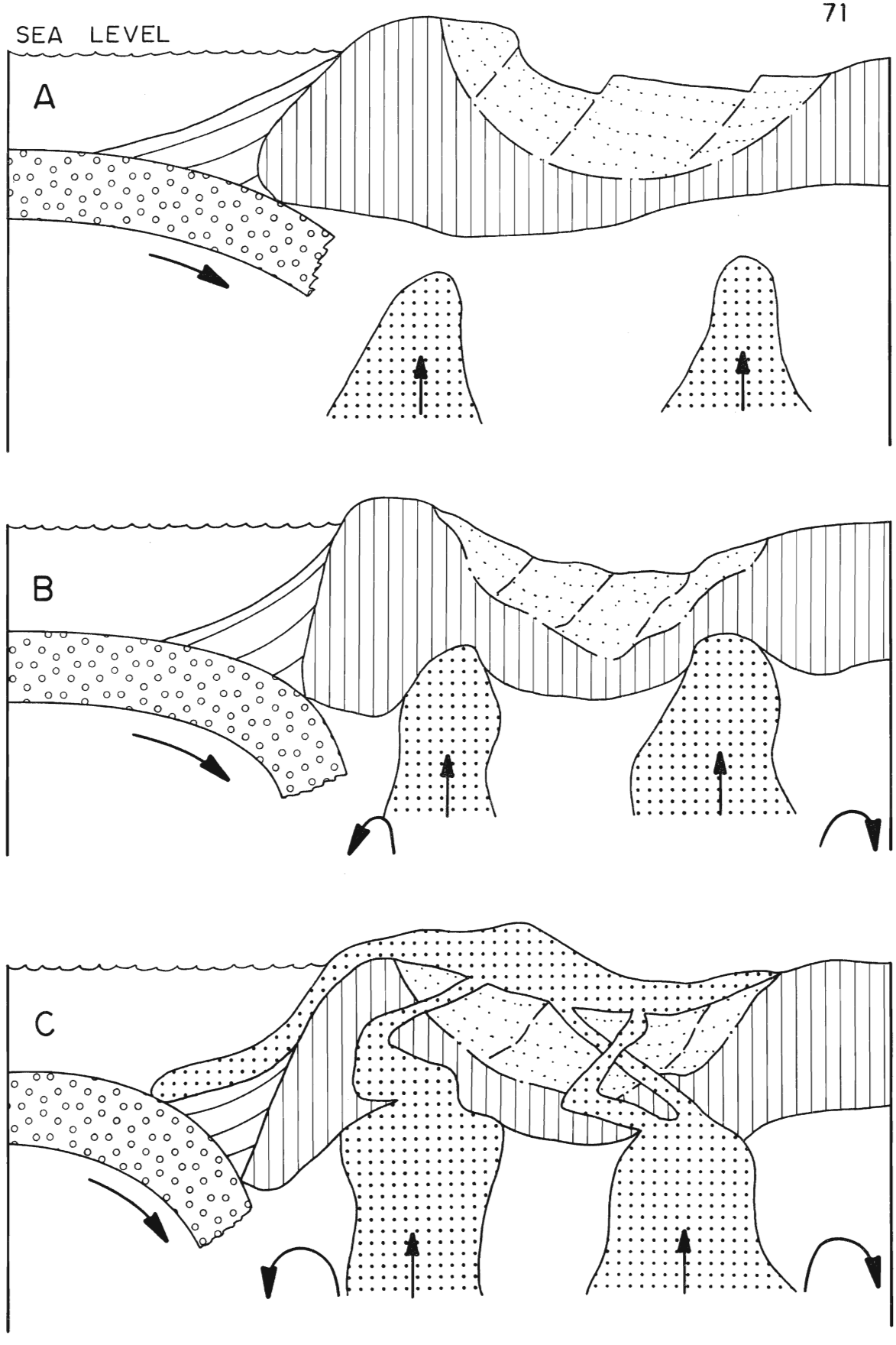

Figure 25.

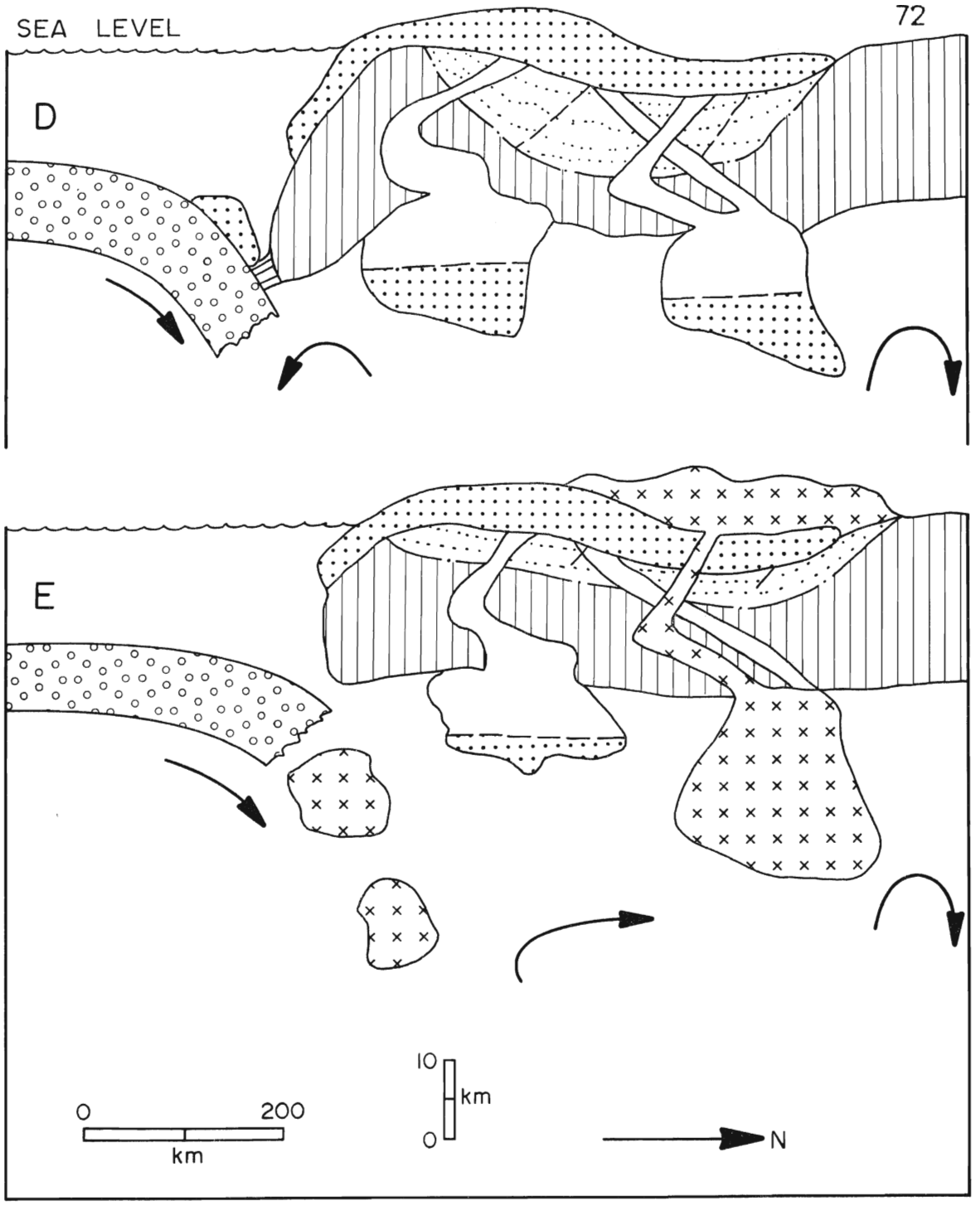

Figure 25. Tectonic model illustrating the evolution of Elliot Lake volcanic rocks (modified after McNaughton 1985).

of basaltic material, as suggested by Condie (1982) which is delivered to the earth's surface through fissures in the crust.

As tholeiitic eruptions become depleted, residual magma in the magma chamber continues fractionating which would account for higher Si, Ba and Zr aconcentrations (Tarney *et al.* 1982). During this time the continental crust is unstable (Sims 1976) and strong Precambrian convection cells in the upper mantle initiate downwarping of water rich continental sialic material (Condie 1982). This Si and H_2O rich slab descends into the upper mantle because it is colder and denser than surrounding mantle material.

Partial melting of this crustal slab occurs and this causes a melt to form which becomes less dense as it is heated and begins to ascend as a diapiric body (Tuttle and Bowen 1958). The ascent of this magma through the upper mantle is aided by circulating convection cell pathways which direct the liquid into previously occupied tholeiitic chambers originally present as a result of rifting. This would account for the high Si and Zr concentrations of these rocks because crustal material is rich is Si and Zr (Pearce 1982) though authors such as Green (1975) and Tarney *et al.* (1982) insist that increasing values of these elements is due to fractionation of parental basaltic material only. The evential eruption of these lavas produce the calc-alkaline lava piles that lie upon the tholeiitic rocks. Overlying the volcanics are various sedimentary units which were derived by abrasion of the greenstone-granite terrain north of the basin (Frarey and Roscoe 1970; Long 1978). The Huronian sequence forms a southerly thickening wedge of platform type sedimentary units at the southern margin of the Southern structural province (Frarey and Roscoe 1970; Long 1978). This sedimentary sequence was dominated by coarse,

cross bedded and fluvial sands that were deposited after downwarps of the basin or unstable platform (Frarey and Roscoe 1970). Two major stream systems were active during deposition of Huronian sediments (Long 1978). One major stream flowed east in the Huronian and Quirke Lake Syncline areas, while another major drainage system was derived in the northeast and crossed the Cobalt plain area. The two streams met in the Huronian area and deposited different clastic material, as granitic material was derived from the western source and granitoid and greenstone fragments were delivered from the northeastern source (Sims 1976; Long 1978). Infilling of the basin was terminated by the Penokean Orogeny (Sims 1976) and during this time folding and metamorphism subjected these rocks to the greenstone metamorphic facies.

CONCLUSIONS

The Elliot Lake region of Ontario contains an assemblage of volcanic and sedimentary rocks that comprise the Huronian Supergroup. The lavas are predominatly composed of secondary mineral assemblages of characteristic of greenschist metamorphic conditions. The temperature of metamorphism ranges between 325°C and 450°C, at pressures between 2.4 and 4.7 kb.

Petrochemical and petrographic evaluation of the volcanic rocks has been carried out using immobile elements such as Ti, Zr, Y and Ni. These elements show changes in concentration that occur in response to fractional crystallization, and are not affected by metamorphic processes. Two volcanic rock associations are distinguished by AFM and ZTN diagrams. Tholeiitic association is characterized by high FeO (averaging 11.5 wt. %), high TiO_2 (averaging 1.62 wt. %) and illustrate a distinct iron and titanium enrichment. Calc-alkaline association illustraates a depletion in FeO and TiO_2 with increased fractional crystallization. FeO and TiO_2 average 6.09 and 1.06 weight percent respectively. Calc-alkaline magmas exhibit an increase in Zr and Si concomitant with the drastic depletion in FeO and TiO_2.

Tholeiitic lavas formed in a continental environment occur as massive accumulations of fissure fed flows. Lavas were produced by partial melting of mantle material and the characteristic iron and titanium enrichment trends reflect fractional crystallization at approximately 10 kb or 30 km under decreasing pO_2 conditions. Calc-alkaline rocks form as a result of fractional crystallization of a basaltic magma at shallow depths in a response to an increase in oxygen pressure, or these rocks may result from the partial melting of descending crustal material. Descending Zr rich

lithospheric slabs lead to replenishment of silica and water because crustal material is silica and water rich, and this results in the increased Si and Zr concentrations observed in calc-alkaline rocks. Both models reflect an increase in oxygen pressure causing the oxidation of iron to occur, thus leading to the characteristic iron depletion trend of calc-alkaline rocks and the formation of hematite. Water enrichment occurs in these magmas in response to increasing oxygen pressure, and this enhances the formation of vesicles.

The Ti-Zr-Ni ternary diagram separates tholeiites from calc-alkaline rocks and reveals the general enrichment of Zr and TiO_2 in Elliot Lake volcanics, and also shows a greater concentration of Ni than for other Thessalon Formation volcanic rocks. Tholeiitic lavas normally reach the earth's surface through fissures developed in the crust.

Tholeiitic lavas begin as partial melts of mantle material and ascend magma chambers as a result of convection cells. Fractionation of this basic liquid leads to iron and titanium enrichment trends.

Calc-alkaline rocks illustrate the development of iron oxides due to an increase in oxygen and water contents of alkaline rich liquids. These lavas result from fractionation of basaltic material or melting of continental crust.

The increased concentrations of Si and Zr indicate that calc-alkaline lavas may be the result of continental crust melting, due to the weight of overlying tholeiitic lavas and downwarping processes initiated by convection cells just under the crust.

Volcanism ceased and was followed by the deposition of Huronian sediments. Sedimentation buried the volcanic rocks and this was

terminated by the Penokean Orogeny which subjected these rocks to greenschist metamorphic conditions.

REFERENCES

Alder, H. L., and Roessler, E.B. 1968. Introduction to Probability and Statistics. W. H. Freeman and Company, San Francisco , 333 pp.

Anderson, A. T., Swihart, G. H., Artolli, G. and Geiger, C.A.1984. Segregation vesicles, gas filter-pressing, and igneous differentiation. Jour. Geology, 92, pp. 55-72.

Arndt, N. T.1977. Ultra basic magmas and high degree melting of the mantle. Contrib. Mineral. Petrol., 64, pp. 205-221.

Arth, J. C., and Hanson, G. N. 1978. Geochemistry and origin of the early Precambrian crust of northeastern Minnesota. Geochim. et Cosmochim. Acta., 39, pp. 352-562.

Bennett, G. 1977. Huronian volcanism, district of Algoma and Sudbury. In Summary of Fieldwork 1977, by the Geological Branch, Ontario Geol. Surv. Misc.Paper 75, pp.102-103.

Bennett, G. 1978. Huronian volcanism, districts of Sault Ste. Marie and Sudbury. In Summary of Fieldwork 1978, by the Geological Branch, Ontario Geol. Surv. Misc. Paper 82, pp.105-110.

Bennett, G. 1982. Geology of the Two Horse Lake area, district of Algoma. Ontario Geol. Surv. Report 210, 63 pp.

Bottrill, T. J. 1971. Uraniferous conglomerates of the Canadian Shield. In Report of Activities, Part A. Geol. Surv. Canada Paper 71-7, 259 pp.

Canon, W. F. 1971. The Penokean Orogeny in northern Michigan. Paper given at Huronian Symposium 1971, annual meting of Geol. Assoc. Can., Sudbury.

Card, K. D., Church, W. R., Franklin, J. M., Frarey, M. J., Robertson, J. A., West, G. F. and Young, G. M. 1977.The Southern Province. In Variations in Tectonic Styles in Canada. Edited by R.A. Price and R. J. W. Douglas. Geol. Assoc. Canada, Special Paper 11, pp. 335-380.

Card, K. D. and Robertson, J. A.1975. North shore of Lake Huron region. Ministry of Natural Resources , Geological Guidebook No. 4.

Card, K. D. and Palonen, P. A. 1976. Geology of the Dunlop-Shakespeare area, district of Sudbury. Ontario Div. Mines G. R. 139, 52 pp.

Card, K. D., Innes, D. G. and Debicki, R. L. 1977.Stratigraphy, sedimentology and petrology of the Huronian Supergroup in the Sudbury-Espanola area. Ont. Dept. Mines, Geosci. Study 16.

Chase, C. G. and Gilmer, T. H. 1973. Precambrian plate tectonics-the midcontinent gravity high. Earth Planet. Sci. Let., 21, pp. 70-78.

Church, W. R. 1971. Comparitive geology of early Proterozoic and early Phanerozoic orogenic belts. Paper given at Huronian Symposium, annual meeting of Geol. Assoc. Canada, Sudbury.

Church, W. R. and Young, G. M. 1970. Disscussion of the progress report of the Federal Provincial Committee on Huronian stratigraphy. Can. J. Earth Sci., 7, No. 3, pp. 912-918.

Clarke, D. B. 1975. Tertiary basalts dredged from Baffin Bay. Can. J. Earth Sci. 12, pp. 1396-1405.

Collins, W. H. 1925. North shore of Lake Huron. Geol. Surv. Canada Memoir 143, 137 pp.

Condie, K. C. 1982. Archean andesites. In Orogenic Andesites and Related Rocks. Edited by R. S. Thorpe, pp. 575-590.

Cox, K. G. and Hornung, G. 1966. The petrology of the karroo basalts of Basutoland. Am. Mineralogist 51, pp. 1414-1432.

Cox, K. G. and Bell, J. D. 1972. A crystal fractionation model for the basaltic rocks of the New Georgia Group, British Islands. Contr. Mineral and Petrol. 37, pp. 1-13.

Cox, K. G., Bell, J. D. and Pankhurst, R. J. 1979. The interpretation of igneous rocks. George Allen and Unwin, Boston, 450 pp.

Davison, J. G. 1984. An Archean Calc-Alkaline Complex, Lake of the Woods, Ontario. Unpub. M.Sc. thesis, Brock University, 303 pp.

Dietrich, R. V. and Skinner, B. I. 1979. Rocks and Rock Minerals. John Wiley and Sons, New York, 319 pp.

Donnelly, T. W. 1963. Genesis of albite in early orogenic volcanic rocks. Am. Jour. Sci., 261, pp. 957-972.

Dostal, J., Coulon, C. and Dupuy, C. 1982. Cainozoic andesitic rocks of Sardina, Italy. In Orogenic Andesites and Related Rocks.Edited by R. S. Thorpe, pp. 353-399.

Eisbacher, G. H. 1969. Regional structural study in the Elliot Lake area, Ontario. In Report of Activities, Part A, April to October 1968. Geol. Surv. Canada 69-1, pp. 158-161.

Fairbairn, H. W., Hurley, P. M., Card K. D. and Knight, C. J. 1969. Correlation of radiometric ages of Nipissing Diabase and Huronian metasediments with Protozoic orogenic events in Ontario. Can. J. Earth Sci., 6, No. 3, pp. 489-497.

Frarey, M. J. 1967. Three new Huronian formational names. Geol. Surv. Canada, Paper 67-6, 3 pp.

Fischer, M. W. 1984. Thrust volcanics in the North Pyrenees. Journal of Structural Geology, 6, No. 6, pp. 721-727.

Flagler, P. 1985. Igneous and Metamorphic Petrography and Geochemistry of the Mafic Huronian Volcanics of the Thessalon Area, Ontario. Unpub. B.Sc. thesis, Brock University, 56 pp.

Frarey, M. J. and Roscoe, S. M. 1970. The Huronian Supergroup north of Lake Huron. In Symposium on Basins and Geosynclines of the Canadian Shield. Edited by A. J. Baer. Geol. Surv. Canada, Paper 70-40, pp. 143-157.

Green, D. H. 1975. Genesis of archean peridotitic magmas and constraints on archean geothermal gradients and tectonics. Geology, 3, pp. 13-18.

Green, D. H., Nicholls, I. A., Viljoen, M. I. and Viljoen, R. D. 1975. Experimental demonstration of the existence of peridotitic liquids in early archean magmatism. Geology, 3, pp. 11-14.

Green, N. L. 1975. Archean glomeroporphyritic basalts. Can. J. Earth Sci., 12, pp. 1770-1784.

Green, T. H. and Ringwood, A. E. 1968. Genesis of the calc-alkaline rock suite. Contr. Mineral. and Petrol., 18, pp. 105-162.

Hamilton, D. L., Burnam, C. W. and Osborn, E. F. 1964. The solubility of water and effects of oxygen fugacity and water content on crystallization in mafic magmas. Jour. Petrol., 5, pp. 21-39.

Harding, W. D. 1941. Geology of the Flack Lake area. Ontario Dept. Mines Report 48,12 pp.

Harris, D. M. 1981. The concentration of CO_2 in submarine tholeiitic basalts Jour. Geology, 89, pp. 689-701.

Henderson, p. 1982. Inorganic Geochemistry. Pergamon Press. Oxford, 353 pp.

Hietanen, A. 1974. Amphibole pairs, epidote minerals, chlorite, plagioclase, in metamorphic rocks, northern Sierra Nevada, California. Am. Miner., 59, pp. 22-40.

Hobbs, B. E., Means, W. D. and Williams, P. F. 1976. An Outline of Structural Geology. John Wiley and Sons, New York, 571 pp.

Holdaway, M.J. 1965. Basic regional metamorphic rocks in part of the Klamath Mountains, northern California. Am. Mineralogist, 50, pp. 953-975.

Holdaway, M. J. 1972. Thermal stability of Al-Fe epidote as a function of fO_2 and Fe content. Contr. Mineral. and Petrol., 37, pp. 307-340.

Hormann, P. K. and Raith, M. 1972. Formation of Al-Fe(III) epidotes. Contr. Mineral. and Petrol., 38, pp. 307-319.

Humphris, S. E. and Thompson, G. 1978. Trace element mobility during hydrothermal alteration of oceanic basalts. Geochim. et Cosmochim. Acta, 42, pp. 127-136.

Huppert, H. E., Sparks, S. J., Turner, J. S. and Arndt, N. T. 1984. Emplacement and cooling of komatiite lavas. Nature, 39, pp. 19-22.

Hurst, J. M. and Surlyk, F. 1982. Depositional environments along a carbonate ramp to slope transition in the Silurian of Washington Land, north Greenland. Can. J. Earth Sci., 20, pp. 473-499.

Hyndeman, D. W. 1972. Petrology of Igneous and Metamorphic Rocks. McGraw-Hill Ltd., New York, 533 pp.

Irvine, T. N. and Baragar, W.R. 1971. A guide to the chemical classification of the volcanic rocks. Can. J. Earth Sci., 8, pp. 523-548.

Jakes, R. and White, J. R. 1972. Major and trace element abundances in volcanic rocks of orogenic areas. Geol. Soc. America Bull., 83, pp. 29-40.

Jamieson, B. G. 1966. Evidence on the evolution of basaltic magma at elevated pressures. Nature, 212, pp. 243-246.

Jolly, W. T. 1970. Zeolite and prehnite-pumpellyite facies in south central Puerto Rico. Mineralogy and Petrology, 27, pp. 204-224.

Jolly, W. T. 1978. Metamorphic history of the Archean Abitibi belt. In Metamorphism in the Canadian Shield. Geol. Surv. Can. Paper 78-10, pp. 63-78.

Jolly, W. T. 1980. Development and degradation of Archean lavas, Abitibi area, Canada, in light of major element geochemistry. Journal of Petrology, 21, No. 2, pp. 323-363.

Jolly, W. T. and Smith, R. E. 1972. Degradation and metamorphic differentiation of the Keweenawan tholeiitic lavas of northern Michigan, U.S.A. Journal of Petrology, 13, pp. 273-309.

Khitarov, N. I. and Kadik, A. A. 1973. Water and carbon dioxide in magmatic melts and peculiarities of the melting process. Contrib. Mineral. and Petrol., 41, pp. 205-215.

Kresz, D. 1984. Geochemistry and Stratigraphy of part of the Western Wabigoon Subprovince in the Kawashegamuk Area, Northwestern Ontario. Unpub. M.Sc. thesis, Brock University, 262 pp.

Kuniyoshi, S. and Liou, J. G. 1974. Contact metamorphism of the Karmutsen volcanics, Vancouver Island, British Columbia. Journal of Petrology, 17, pp. 73-99.

Levi, B. 1969. Burial metamorphism of a Cretaceous volcanic sequence west from Santiago, Chile. Contr. Mineral. and Petrol., 24, pp. 30-49.

Liou, J. G. Kiuniyoshi, S. and Ito, K. 1974. Experimental studies of the phase relations between greenschist and amphibolite in a basaltic system. Am. Jour. Sci., 274. pp. 613-632.

Long, D.F. 1977. Resedimented conglomerates of Huronian ages from the north shore of Lake Huron, Ontario, Canada. Can. J. Earth Sci., 14, pp. 2495-2509.

Long, D. F. 1978. Depositional environments of a thick Proterozoic sandstone.The Mississagi Formation of Ontario, Canada. Can. J. Earth Sci., 15, pp. 190-206.

Long, D. F. 1981. The sedimentary framework of placer gold concentrations in basal Huronian strata of the Cobalt embayment. Ontario Geological Survey M. P. 100, pp. 218-223.

Lumbers, S. B. 1971. Geology of the North Bay area, districts of Nipissing and Parry Sound. Ontario Div. Mines, G. R. 94, 104 pp.

MacLean, W. H., Seymour, K. and Prabhu, M. K. 1982. Sr, Y, Zr, Nb, Ti and REE in Grenville amphibolites at Montaubanles-Mines, Quebec. Can. J. Earth Sci., 19, pp. 633-644.

MacPherson, G.J. 1984. A model for predicting the volume of vesicles in submarine basalts. Jour. Geology, 92, pp.73-82.

McNaughton, J. 1985. A Petrological and Geochemical Comparison of the Felsic and Mafic Volcanics of the Thessalon Area, Ontario. Unpub. B.Sc. thesis, Brock University, 64 pp.

Meyn, H. D. 1973. The proterozoic sedimentary rocks north and northeast of Sudbury, Ontario. In Huronian Stratigraphy and Sedimentation. Geological Association of Canada, Special Paper 12.

Miyashiro, A. 1973. Metamorphism and Metamorphic Belts. George Allen and Unwin Ltd., London, 892 pp.

Miyashiro, A. and Seki, Y. 1958. Enlargement of the composition field of epidote and piemontite with rising temperature. Amer. Jour. Sci., 256, pp. 423-430.

Morrison, M. A. 1978. The use of immobile trace elements to distinguish the paleotectonic affinities of metabasalts. Applications to the Paleocene basalts of Mull and Skye, northwest Scotland. Earth Planet Sci. Let., 39, pp. 407-416.

Moore, J. G. 1979. Vesicularity and CO_2 in mid-ocean ridge basalt. Nature, 282, pp. 250-253.

Muenow, D. W., Graham, D. G., Liu, N. W. and Delaney, J. R. 1979. The abundance of volatiles in Hawaiian tholeiitic submarine basalts. Earth Planet. Sci. Let., 42, pp. 71-76.

Myer, G. H. 1966. New data on zoisite and epidote. Am. Jour. Sci., 264, pp. 364-385.

Nesbitt, R. W. and Sun, S. 1976. Geochemistry of Archean spinifex-textured peridotites and magnesium and low magnesium tholeiites. Earth Planet. Sci. Let., 31, pp. 433-453.

Nisbet, E. G. 1984. Turbulence in petrology-the behaviour of komatiites. Nature, 309, pp. 14-15.

Nisbet, E. G. and Walker, D. 1982. Komatiites and the structure of the Archean mantle. Earth Planet. Sci. Let., 60, pp. 105-113.

Osborn, E. F. 1959. Role of oxygen pressure in the crystallization and differentiation of basaltic magma. Am. Jour. Sci., 257, pp. 609-647.

Osborn, E. F. 1962. Reaction series for subalkaline igneous rocks based on different oxygen pressure conditions. Am. Mineral., 47, pp. 211-226.

Pearce, J. A. 1982. Trace element characteristics of lavas from destructive plate boundaries. In Orogenic Andesites and Related Rocks. Edited by R. S. Thorpe, pp. 525-546.

Pearce, J. A. and Norry, M. J. 1979. Petrogenetic implication of Ti, Y, Zr and Nb variations in volcanic rocks. Contrib. Mineral. and Petrol., 69, pp. 33-47.

Pettijohn, F. J. 1970. Symposium on Basins and Geosynclines of the Canadian Shield. Edited by A. J. Baer. Geol. Surv. Canada, Paper 70-40, pp. 262-265.

Pienaar, P. J. 1963. Stratigraphy, petrology, and genesis of the Elliot Lake Group, Blind River, Ontario, including the uraniferoous conglomerate. Geol. Surv. Canada Bull. 83, 143 pp.

Philpotts, J. A., Schnetzler, C. C. and Hart, S. R. 1969. Submarine basalts: Some K, Rb, Sr, Ba, rare-earth, H_2O and CO_2 data bearing on their alteration, modification by plagioclase, and possible source materials. Earth Planet. Sci. Let., 7, pp. 293-299.

Remington,R. D. and Shork, M. A. 1970. Statistics with Applications to the Biological and Health Sciences. Prentice-Hall, Inc., New Jersey,

Ringwood, A. E. 1974. The petrological evolution of island systems. J. Geol. Soc. Lond., 130, pp. 183-204.

Robertson, J. A. 1957. Joubin Township, Algoma district. Ont. Dept. Mines Map 2001.

Robertson, J. A. 1959. Beange Township, Algoma district. Ont. Dept. Mines Map 2015.

Robertson, J. A. 1961. Recent geological investigations in the Elliot Lake-Blind River uranium area, Ontario. Ont. Dept. Mines M. P. 9, 31 pp.

Robertson, J. A. 1961. Geology of townships 143 and 144, district of Algoma. Ont. Dept. Mines Geol. Rept. 4, 65 pp.

Robertson, J. A. 1963. Bouck Township, Algoma district. Ont. Dept. Mines Map 2114.

Robertson, J. A. 1963. Geology of townships 155, 156, 161, 162 and parts of 167 and 168, district of Algoma. Ont. Dept. Mines Geol. Rept. 13, 88 pp.

Robertson, J. A. 1969. Nicholas and Raimbault townships, Algoma district. Ont. Div. Mines Map 2347.

Robertson, J. A. 1971. A review of recently acquired geological data, Blind River-Elliot Lake area. Paper given at Huronian Symposium,annual meeting of Geol. Assoc. Can., Sudbury.

Robertson, J. A. 1971. A review of recently acquired geological data, Blind River-Elliot Lake area. Ont. Dept. Mines M. P. 45, 35 pp.

Robertson, J. A. 1976. Geology of the Massey area, districts of Algoma, Manitoulin and Sudbury. Ont. Div. Mines G. R. 136, 130 pp.

Robertson, J. A., Giblin, P.E. and Leahy, E. J. 1977. Blind River-Elliot Lake sheet, districts of Algoma and Sudbury. Ont. Geol. Surv. Map p. 304.

Roscoe, S. M. 1973. The Huronian Supergroup. A paleoaphebian succession showing evidence of atmospheric evolution in Huronian stratigraphy and sedimentation. Geol. Assoc. Canada, Special Paper 12.

Schiffman, P. and Liou, J. G. 1980. Synthesis and stability relation of Mg-Al pumpellyite. Journal of Petrology, 21, pp. 441-474.

Schuiling, R. D. and Vink, B. W. 1967. Stability relations of some titanium minerals. Geochim. et Cosmochim. Acta, 31, No. 12, pp. 2399-2411.

Sims, P. K. 1976. Precambrian tectonics and mineral deposits, Lake Superior region. Economic Geol., 71, pp. 1092-1127.

Smith, F. G. 1963. Physical Geochemistry. Addison-Wesley Publishing Co., Inc., Mass., 624 pp.

Smith, R. E. 1968. Redistribution of major elements in the alteration of some basic lavas during burial metamorphism. Journal of Petrology, 9, pp. 191-219.

Smith, R. E. 1969. Zones of progressive regional burial metamorphism in part of the Tasman Geosyncline, eastern Australia. Journal of Petrology, 10, pp. 144-163.

Smith, R. E. and Smith, S. E. 1976. Comments on the use of Ti, Zr, Y, Sr, K, P and Nb in classification of basaltic magmas. Earth Planet. Sci. Let., 32, pp. 114-120.

Stout, M. Z. and Nicholls, J. 1977. Mineralogy and petrology of Quaternary lavas from the Snake River Plain, Idaho. Can. J. Earth Sci., 14, pp. 2140-2156.

Strens, R. G. 1963. Some relationships between members of the epidote group. Nature, 198, pp. 80-81.

Studemeister, P. A. 1983. The greenschist facies of an Archean assemblage near Wawa, Ontario. Can. J.Earth Sci., 20, pp. 1409-1420.

Suk, M. 1983. Petrology of Metamorphic Rocks. Elsevier Scientific Publishing Co., New York, 322 pp.

Symons, D. T. 1975. Huronian glaciation and polar wander fron the Gowganda Formation, Ontario. Geology, 3, pp. 303-306.

Tabatabai, M. 1978. Petrography and Geochemistry of McArthur Township Area, Ontario. Unpub. M.Sc. thesis, Brock University, 141 pp.

Tarney, J., Weaver, S. D., Saunders, A. D., Pankhurst, R. J. and Baker, P. F. 1982. Volcanic evolution of the northern Antarctic Peninsula and the Scotia Arc. In Orogenic Andesites and Related Rocks. Edited by R. S. Thorpe, PP. 371-400.

Turner, F. J. 1968. Metamorphic Petrology, Mineralogical and Field Aspects. McGraw-Hill Bood Co., New York, 739 pp.

Tuttle, O. F. and Bowen, N. L. 1958. Origin of granite in light of experimental studies in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O. Geol. Soc. Am. Mem. 74, 153 pp.

Wedepohl, K. H., Correns, C. W., Shaw, D. M., Turekian, K. K. and Zeman, J. 1969. Handbook of Geochemistry. Springer-Verlag, New York, pt. 1.

Wedepohl, K. H., Correns, C. W., Shaw, D. M., Turekian, K. K. and Zeman, J. 1970. Handbook of Geochemistry. Springer-Verlag, New York, pt. 2.

Weichelt, K. 1984. Geochemical and Petrological Analysis of the Thessalon Volcanic Unit, Thessalon, Ontario. Unpub. B. Sc. thesis, Brock University, 41 pp.

Windley, B. F. 1979. The Evolving Continents. John Wiley and Sons, New York, 385 pp.

Winkler, H. 1979. Petrogenesis of Metamorphic Rocks, 5th Edt. Springer-Verlag, New York, 348 pp.

Wood, D. A., Joron, J. L. and Trevil, M. 1979. A re-appraisal of the use of trace elements to classify and discriminate between magma series erupted in dfferent tectonic settings. Earth Planet. Sci. Let., 45, pp. 326-336.

Young, G. M. 1971. Stratigraphic and sedimentological framework of the Huronian rocks of the Southern Province of the Canadian Shield. Paper given at Huronian symposium, annual meeting of Geol. Assoc. Can., Sudbury.

Young, G. M. 1973. Origin of carbonate rich early Proterozoic Espanola Formation, Ontario, Canada. Geol. Soc. America Bull., 84, pp. 135-160.

Young, G. M. 1981. Upper Proterozoic supra-crustal rocks of North America. Precambrian Research,15, pp. 305-330.

APPENDIX I

Major Element Geochemical Data In Weight Percent

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O_5	TOTAL	LOI	TOTAL+LOI
B19	47.56	14.49	16.73	2.75	5.11	4.06	1.59	1.99	0.11	0.34	94.71	4.51	99.22
B13	50.71	14.87	17.70	2.57	2.15	3.64	2.01	2.03	0.12	0.30	96.10	1.86	97.76
B7	44.83	17.37	20.10	2.93	1.49	4.00	1.19	2.52	0.15	0.45	95.04	3.40	98.44
E2	50.67	14.80	12.39	6.65	5.63	3.15	1.42	1.18	0.20	0.07	96.15	2.61	98.76
C1	60.47	14.04	7.43	1.45	4.04	5.08	1.08	1.38	0.09	0.17	95.22	3.67	98.89
A6	49.37	13.68	12.83	6.70	11.36	1.65	0.34	0.99	0.18	0.06	97.16	0.14	97.30
E102	47.36	15.01	18.64	4.03	3.58	3.64	0.34	1.77	0.15	0.29	94.82	4.96	99.78
D5	50.98	13.94	10.96	6.67	9.76	1.50	1.07	0.60	0.18	0.03	95.69	2.29	97.98
A10	52.46	10.98	14.33	6.48	6.16	2.75	0.47	1.73	0.20	0.19	95.74	2.26	98.00
D3	51.39	13.72	11.03	6.55	9.73	1.68	0.92	0.64	0.18	0.05	95.90	1.89	97.79
B15	48.94	15.02	9.50	2.87	7.89	4.93	0.72	1.45	0.20	0.19	91.69	6.83	98.52
B4	48.95	15.77	16.05	4.30	1.66	3.07	1.87	2.13	0.12	0.31	94.22	3.67	97.89
A2	49.90	12.27	22.20	4.13	1.64	1.97	1.06	1.89	0.12	0.18	95.37	3.42	98.79
D4	51.20	13.76	10.67	6.55	10.07	1.41	0.82	0.63	0.18	0.05	95.34	1.96	97.30
B12	46.66	17.21	17.64	4.06	0.62	3.15	2.21	1.75	0.18	0.24	93.73	3.81	97.54
A15	59.57	14.83	8.25	2.98	5.09	4.45	0.53	0.79	0.11	0.15	96.75	1.16	97.91
A18	49.99	12.99	17.46	3.54	3.48	4.41	2.25	2.04	0.17	0.25	96.59	1.24	97.83

Major Element Geochemical Data In Weight Percent

Sample	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	P_2O5	TOTAL	LOL	TOTAL+LOI
C2	56.77	14.11	13.01	4.92	1.01	0.00	3.52	0.81	0.11	0.26	94.52	4.35	98.87
A25	49.70	10.37	15.40	8.16	7.18	3.28	0.43	1.41	0.21	0.14	96.29	3.08	99.37
E1	51.18	12.32	13.52	5.42	9.03	2.09	1.34	1.06	0.19	0.08	96.22	1.71	97.93
01	58.85	14.21	11.11	2.34	0.73	4.41	1.87	1.36	0.06	0.16	95.09	2.23	97.32
B3	44.56	16.39	17.51	4.38	0.57	0.53	8.20	1.79	0.13	0.33	94.38	3.38	97.76
C4	66.56	13.04	6.52	2.04	0.20	0.61	5.24	0.40	0.05	0.06	94.73	2.72	97.45
D9	49.63	13.81	15.04	3.70	4.91	4.41	0.58	1.95	0.20	0.33	94.55	3.20	97.75
A11	49.75	13.50	16.04	5.56	3.01	3.56	1.55	2.09	0.16	0.29	95.51	2.69	98.20
A2	49.50	11.63	22.48	4.30	1.64	2.20	1.04	1.88	0.13	0.19	94.99	3.42	98.41
A182	50.65	12.62	17.97	3.86	3.53	3.79	2.23	2.06	0.16	0.26	97.16	1.24	98.38
A72	48.76	12.35	15.04	5.63	8.76	1.49	1.93	1.57	0.18	0.14	95.85	2.40	98.25
A1	47.72	13.12	15.99	6.09	8.86	1.98	0.42	1.52	0.20	0.11	96.00	2.60	98.60

APPENDIX II

Trace Element Geochemical Data In Parts Per Million

Sample	Ba	Zr	Sr	Y	S	Zn	Cu	Ni	Rb	Ce
B19	521.0	146.2	653.4	8.1	115.0	41.3	41.5	57.4	32.9	48.0
B13	8.10.8	139.6	664.9	8.2	77.0	49.4	0.0	64.3	47.7	46.8
B7	553.0	165.7	649.3	7.4	95.8	59.6	27.5	82.8	18.6	58.2
E2	565.1	85.8	662.9	9.6	72.0	111.1	17.3	130.1	42.0	80.2
C1	589.8	141.5	660.2	12.8	58.1	57.0	104.5	106.9	55.6	69.6
A6	357.8	69.1	624.3	13.8	1711.5	54.6	105.5	68.3	14.3	51.1
E10	344.1	149.1	647.3	8.8	61.2	53.2	343.0	24.5	24.6	51.1
D5	386.5	81.5	661.6	12.2	174.3	72.9	16.7	34.6	52.8	57.2
B4	696.6	155.3	665.4	7.5	42.1	65.6	3.9	214.6	31.4	52.4
A2	367.6	129.1	646.8	4.5	59.9	73.6	52.9	196.0	36.2	40.7
D4	570.4	74.0	656.2	12.3	256.5	88.0	109.0	120.8	37.3	76.5
B12	1381.9	115.3	659.2	7.7	112.4	109.6	0.0	136.2	32.6	71.9
A16	517.3	181.8	666.2	25.4	226.0	116.7	25.9	33.8	29.3	68.1
A18	951.2	155.4	659.2	10.1	142.8	136.5	314.5	85.4	53.7	42.9
C2	847.9	106.2	599.4	6.5	98.9	58.0	0.0	147.4	143.9	75.9
E1	386.5	81.5	661.6	12.2	174.3	72.9	16.7	34.6	52.8	57.2
01	1377.4	122.6	666.0	11.6	37.9	42.6	0.0	124.4	47.0	60.3

Trace Element Geochemical Data In Parts Per Million

Sample	Ba	Zr	Sr	Y	S	Zn	Cu	Ni	Rb	Ce
B3	2691.6	126.5	665.3	5.1	81.3	47.9	17.2	176.2	73.4	50.5
C4	1549.6	163.5	576.0	5.9	42.1	46.1	2.2	27.7	183.0	79.9
D9	502.2	164.5	660.1	17.0	440.9	127.7	18.0	87.0	34.4	50.1
A11	494.3	188.6	666.1	8.6	78.2	164.0	83.6	111.1	50.9	39.6
A2	367.6	129.1	646.8	4.5	59.9	73.6	52.9	196.0	36.2	40.7
A18	951.2	155.4	659.2	10.1	142.8	136.5	314.5	85.4	53.7	42.9
A72	395.8	91.7	653.1	17.4	2726.3	60.2	80.7	54.9	63.1	53.7
A1	324.3	83.4	651.9	14.1	222.1	85.9	18.6	46.3	8.3	48.4

APPENDIX III

Statistical Data

Sample	FeO/MgO(X)	FeO(X)	TiO_2(X)
B19	4.29	11.89	2.01
B13	4.93	12.91	2.08
B7	4.90	14.57	2.56
E2	1.30	8.77	1.20
C1	2.80	4.10	1.40
A6	1.38	9.50	1.02
E10	3.40	13.71	1.78
D5	1.19	8.07	0.62
A10	1.53	10.12	1.76
D3	1.22	8.18	0.65
B15	2.04	5.95	1.47
B4	2.58	11.33	2.18
A2	4.05	16.97	1.91
D4	1.17	7.89	0.65
B12	3.16	13.19	1.79
A15	1.79	5.45	0.81
A18	3.52	12.69	2.09
C2	2.40	11.97	0.82
A25	1.72	14.12	1.42
E1	2.27	12.58	1.08
A1	2.39	14.77	1.55
01	4.37	10.49	1.40
B3	3.62	16.21	1.84
C4	2.93	6.16	0.41
D9	3.73	14.10	2.00
A11	2.63	14.94	2.13
A2	4.74	20.71	1.92
A182	4.26	16.70	2.10
A7	2.43	13.96	1.60

APPENDIX IV

Accuracy of major oxides and trace element of standard rocks analysis based on least squares regression taken from Payne (1978)

Oxides	Wt.%	Trace Elements	ppm
SiO_2	+/- 1.00	Ba	+/- 129.73
Al_2O_3	+/- 1.00	Zr	+/- 47.76
FeO	+/- 0.20	Sr	+/- 50.12
MgO	+/- 0.05	Y	+/- 9.33
CaO	+/- 0.20	Zn	+/- 14.39
Na_2O	+/- 0.30	Cu	+/- 8.66
K_2O	+/- 0.05	Ni	+/- 10.79
TiO_2	+/- 0.01	Rb	+/- 18.40
MnO	+/- 0.05		
P_2O_5	+/- 0.05		

APPENDIX V

Sample Localities

Sample	Longitude (W)	Latitude (N)
B19	82°40'	46°30.7'
B13	82°40.2'	46°30.6'
B7	82°39.8'	46°30.8'
E2	82°54.5'	46°34.5'
C1	82°40.5'	46°30.8'
A6	82°31.5'	46°23.5'
E10	82°54.5'	46°34.5'
D5	82°47'	46°32'
A10	82°30.5'	46°32'
D3	82°47'	46°32'
B15	82°40.2'	46°30.6'
B4	82°34.5'	46°30.8'
A2	82°32'	46°23.5'
D4	82°47.5'	46°32'
B12	82°40.2'	46°30.6'
A15	82°30'	46°23'
A18	82°29.5'	46°23'
C2	82°40.5'	46°30.8'
A25	82°29'	46°23'
E1	82°54'	46°34'
A1	82°32'	46°23.5'
01	82°40'	46°30.8'

B3	82°39.5'	46°30.5'
C4	82°40.8'	46°30.8'
D9	82°46.5'	46°31.5'
A11	82°30.5'	46°23.5'
A2	82°32'	46°23.5'
A182	82°29.5'	46°23'
A7	82°31.5'	46°23.5'